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CONTENTS

Distribution of mountain building in geologic time. By James Gilluly.... 561–590

Karroo dolerites of the Union of South Africa. By Frederick Walker and Arie Poldervaart 591–706

Lower Middle Ordovician of south-central Pennsylvania. By Lawrence C. Craig 707–780

Portion of the San Andreas rift in southern California. By Robert E. Wallace 781–806

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PAPERS IN PRESS FOR FORTHCOMING ISSUES

STRUCTURE OF THE NORTHERN MUDDY MOUNTAIN AREA, NEVADA. By Chester R. Longwell
This paper contains a colored geologic map of the area (20 by 18 inches).
Scale 1 inch = 1 mile. Separate copies of the map (unfolded, in tubes) may be purchased.
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SUBSURFACE CONSTITUTION OF BIKINI ATOLL AS INDICATED BY A SEISMIC REFRACTION SURVEY. By M. B. Dobrin, Beauregard Perkins, and B. L. Snavely.

OBSERVATIONS IN SOME YELLOWSTONE GEYSERS. By F. Donald Bloss and Tom. F. W. Barth

UPPER PLIOCENE VERTEBRATES FROM KEEFE CANYON, MEADE COUNTY, KANSAS. By C. W. Hibbard and Elmer S. Riggs

GEOLOGIC AND GEOGRAPHIC RECONNAISSANCE OF EASTERN MARKAGUNT PLATEAU, UTAH. By Herbert E. Gregory

STRUCTURE AT KIRKLAND LAKE, ONTARIO, CANADA. By Harold Hopkins

STRATIGRAPHY AND ORIGIN OF BAUXITE DEPOSITS. By E. C. Harder

INTRUSIVES OF CENTRAL MINNESOTA. By Margaret Skillman Woyski

PETROLOGY OF VOLCANIC ROCKS OF NORTHEASTERN NEW MEXICO. By Helen Stobbe

VOLCANIC ROCKS OF NORTHEASTERN NEW MEXICO. By R. F. Collins

BULLETIN OF THE GEOLOGICAL SOCIETY OF AMERICA
VOL. 60, PP. 561-590, 16 FIGS. APRIL 1949

DISTRIBUTION OF MOUNTAIN BUILDING IN GEOLOGIC TIME

BY JAMES GILLULY

(*Presidential Address of the Retiring President of The Geological Society of America*)

The title best descriptive of this paper has already been used by Shepard (1923) 25 years ago: *To question the theory of periodic diastrophism.* As my purpose is to question the theory, I will state at the outset my understanding of it, using as my text the statements of Stille (1924; 1936), perhaps the leading present-day exponent.

This theory holds that most of geologic time has been anorogenic; that is, the crust of the earth has been generally stable, though always subject to broad regional uplifts, depressions, and warpings—the so-called epeirogenic movements. From time to time this ruling quiet has been punctuated by geologically very short orogenic epochs. During these episodes of storm and revolution, parts of the crust widely scattered over the earth have been simultaneously strongly deformed, folded, and sheared. Though no unconformity can be really world wide, the unconformities that do occur to mark these orogenic periods, whether in the Himalayas, South Africa, western Europe, or North America, will be found limited to the same small segments of the geologic record. Any particular unconformity may be absent from one or several of the other areas, but if any is found in them it will be within the same narrow time range.

In 1924 Stille listed about 30 such orogenic episodes, each represented in more than one area, since the beginning of the Cambrian. In 1936, after he had visited America, his revised list had grown to about 40 such epochs, though each orogeny by hypothesis is still world wide. He estimated the duration of each such epoch as about 300,000 years. Thus the 40 orogenies would require only about 12 million years. This is less than 2½ per cent of the time elapsed since the beginning of the Cambrian, according to the best data from radioactivity (Holmes, 1947). If this thesis can be established, the basic tenets of the periodicists are, of course, well founded.

According to this theory, the present mountainous condition of the earth is abnormal. We are now in the dying phase of an orogenic period or have actually entered upon an anorogenic one, during the course of which the present mountains will be carved away, and the relief of most of the earth will be reduced to approach that of the Mississippi Valley before, some millions of years hence, we enter a new period of world-wide orogeny.

Practically this same view has prevailed for many years. In this or closely similar form it has had the support of T. C. Chamberlin (1909), Barrell (1917), Kuenen (1941), Umbgrove (1947), Holmes (1945), Bucher (1933), and R. T. Chamberlin (1944). Not only have these geologists accepted this theory, but some have held that periodicity is so well established as to furnish the fundamental basis for stratigraphic correlation. Several have also held that orogenic activity is speeding up with time, as proved by the greater stratigraphic thicknesses per million years of time span of the younger systems and series as compared with the older. Now these

authorities are very weighty, and geophysicists such as Joly (1925), Griggs (1939), and Vening-Meinesz (1934, 1948) have followed their lead and devised mechanisms designed to yield periodic diastrophism—even though they have all had trouble in finding any accelerating mechanism. However, if the periodicity is not real, and mountain-making movements are going on at one place or another all the time, such theories would have to be modified or abandoned.

On this subject, as on most others in our science, there have always been some highly respectable skeptics. Shepard (1923), Berry (1929), Von Bubnoff (1931), Arkell (1933), Woodring (1938), and Spieker (1946), among this group, come readily to mind. Nevertheless, the theory of periodicity, with its corollary of increasing tempo with time, seems clearly dominant. I have already mentioned that the rest of my paper is devoted to questioning the grounds on which it rests. I am not ready to assert that orogeny is actually continuous but I do believe that none of the arguments yet advanced in favor of periodicity is strong, most are very weak, and some, indeed, amount to prejudgements.

I wish to discuss five main aspects of the problem: (1) the present tectonic condition of the earth; (2) the significance of the Tertiary record; (3) the significance of the volumes of sedimentary rocks to the relief of the continents; (4) the imperfections of the geologic record; and (5) the precision of geologic dating.

First, is the present an anorogenic period? Let us examine the record from California.

The Long Beach earthquake of March 10, 1933, was a moderate shock, but because it originated near the city it was enormously destructive of the flimsy buildings there. Benioff (1938) concluded from the seismograph records that the shock was caused by slip along a northwesterly trending fault, offshore between Newport and Long Beach. The U. S. Coast and Geodetic Survey had run precise levels in the area in 1931; these were repeated a few months after the quake (U. S. Coast and Geodetic Survey, 1941). Comparison of these surveys reveals that an area in the Alamitos Plain east of Long Beach had been bowed upward—presumably at the time of the quake—into a gentle arch about 7 inches high and 4 miles across. As the southerly part of the plain, nearer the epicenter, was not surveyed, the uplift there may have been greater (Fig. 1). The changes in elevation are too systematic to be explained by lurching of surficial materials; the upwarping must indicate anticlinal folding at depth.

Grant and Sheppard (1939) have analyzed repeated level data from the Inglewood area, of the Los Angeles plain, and have shown that the northwestern part of the Baldwin Hills is rising at the rate of 3 feet per century, though no strong local earthquakes have been recorded during the 5 years covered by their study.

The Coast Survey ran precise levels across Cajon Pass, north of San Bernardino, California, in 1906, 1924, and 1944 (R. D. Horne, personal communication, 1946). These surveys prove that the country between San Bernardino and Victorville, across the pass in the Mohave Desert, has been arched upward (Fig. 2). The crest of this arch lies north of the San Andreas rift, and the level data show no discontinuity at the fault, which has been inactive in this segment during the time span involved. Again the arching is not explicable by surficial movements; it must reflect upwarping

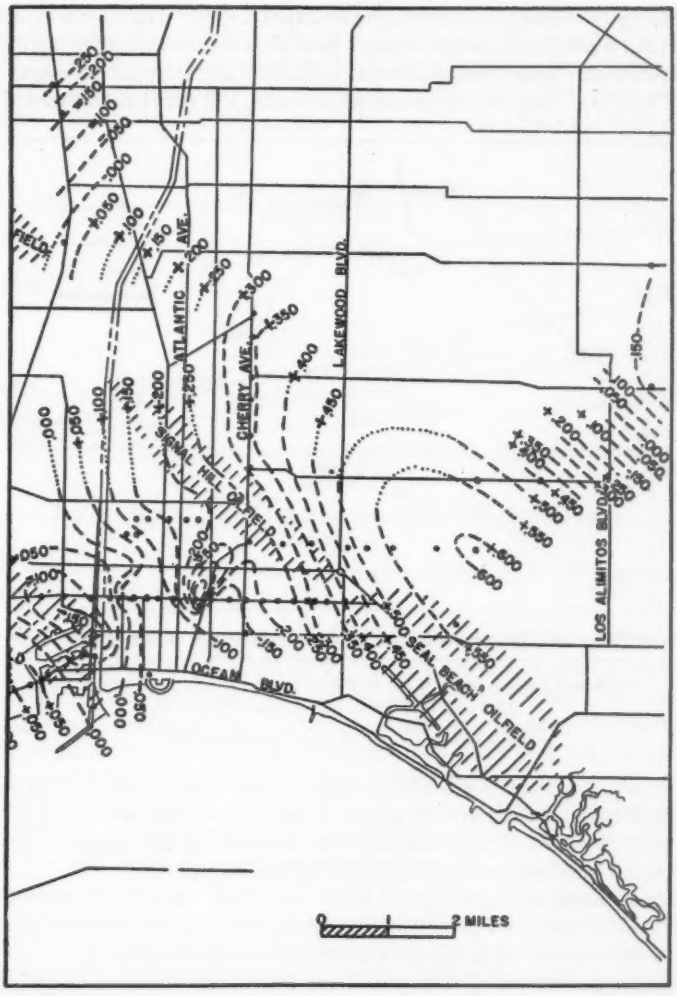

FIGURE 1.—*Uplift at Long Beach, California, shown by comparison of leveling of 1931 and 1933–1934*

U. S. Coast and Geodetic Survey, precise levels. The Long Beach earthquake of March 10, 1933, took place in the interval between the surveys.

of deeper rocks. During the 38 years between the earliest and the latest surveys the crest of the arch has risen .195 meter (nearly 8 inches) or at the rate of nearly 20 inches per century.

In the Buena Vista oil field in the southern San Joaquin Valley, a thrust fault was found in 1933, advertising itself forcefully by cutting off several oil-well casings. The points of failure of the casings fall in a plane that dips 25°NE. and emerges at the

surface along a low scarp that had hitherto escaped notice. That this is a fault scarp became clear when pipe lines laid across it in shallow trenches began to buckle out of the ground, power lines crossing it began to slacken, and the road pavement to rupture. There have been no perceptible earthquakes, but the fault has moved at the

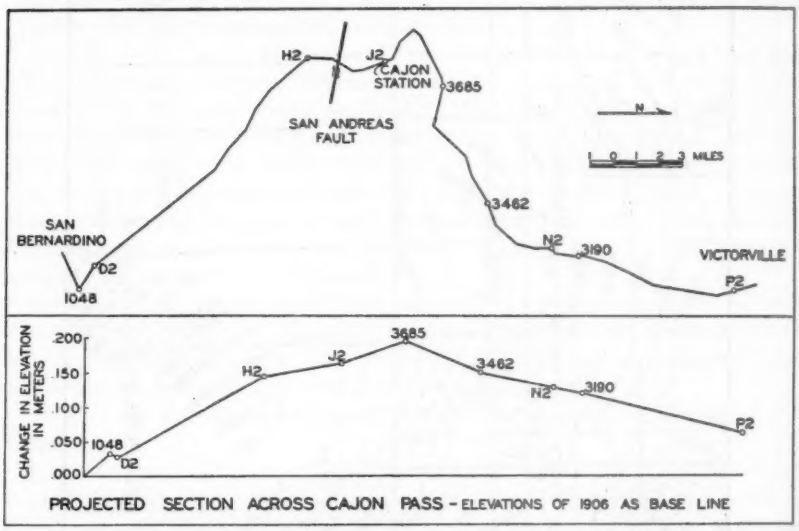

FIGURE 2.—*Differences between levels of the several reoccupied bench marks shown by the surveys of 1906 and 1944 across Cajon Pass, California*

Precise level data from U. S. Coast and Geodetic Survey.

rate of about an inch a year since it was first noticed. Since 1933 the northeastern block has risen about 7 inches with respect to the southwestern, and the span between power poles anchored on opposite sides has shortened about a foot. The rate of uplift is about 4 feet per century (Koch, 1933; McMasters, 1943).

By human standards these rates of uplift are trivial. But if they continue for a time as long as that since the end of the Pleistocene, say about 25,000 years, they would notably modify the topography of southern California (Fig. 3).

We cannot predict the next fault at Long Beach, but the earthquake history of California suggests one or two more in the next century. If these rates continued for 25,000 years, the Alamitos Plain would be a hill about 300 feet high. If Cajon Pass is a fair sample of the San Bernardino-San Gabriel ranges these mountains would then be 400 feet higher than now. The Buena Vista Hills would have grown about 1000 feet. We will have to correct these uplifts for erosion, of course, but it should be emphasized that, even if erosion wiped them out, the structure—which is our main record of orogeny—would remain. That erosion would not, however, obliterate them seems likely from the studies of denudation in southern California by the U. S. Forest Service. These data suggest a correction of less than 30 per cent of these

figures even for mountains carved from soft Tertiary rocks; the correction would average much less and for the crystalline of the San Gabriel Range probably would be only 5 to 10 per cent (Unpublished data, U. S. Army Engineers, Los Angeles Office). If we dare to project these rates for 200,000 years (still only a fifth of the usual estimate of the duration of the Pleistocene and only two-thirds of one of Stille's orogenic

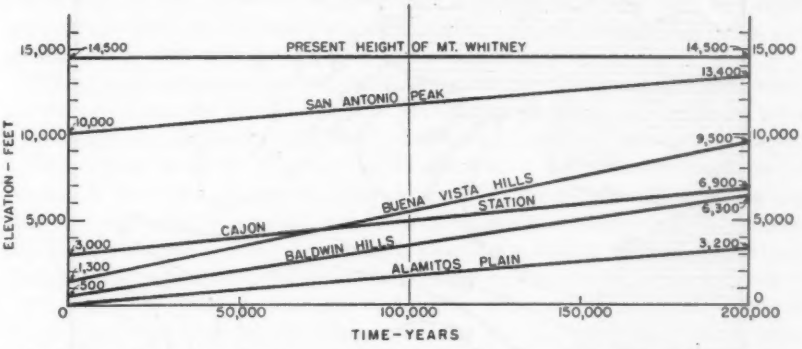

FIGURE 3.—*Altitudes in areas of known uplift in California, projected into the future at presently indicated rates*

Intended merely to show the adequacy of present movements to form mountains—not that such mountains will necessarily develop.

periods) the peaks of the San Gabriel would be approaching the height of Mount Whitney, the Alamitos Plain would be higher than the Santa Monica Mountains (about 2500 feet), and the Buena Vista Hills would be about as high as the Cascades.

Though I am a loyal Huttonian—as will be abundantly evident before I conclude—and believe that the present is the key to the past, I am not so reckless as to prophesy that these movements will continue. But I think they prove that the crust in southern California is far from quiet. They are random samples that happened to be noticed because of precise levels or oil-field work. There is no reason to think these are the most extreme movements actually going on. We are living in a time of active movement that I venture to think is a typical orogenic epoch. These measured rates could account for any mountains on earth in a geologically reasonable time. To make a Mount Everest would take about 2 million years—certainly not an unreasonably high estimate of the time actually involved in raising the Himalayas. The coarse Siwalik beds that range throughout the upper Cenozoic of the foothills suggest the actual uplift took 10 times that long. Comparable rates of movement have been measured in many areas of Japan. Tsuboi (1939) reports consistent uplift of the Idu Peninsula at variable rates as high as 1 mm. in 2 days. Some bench marks rose 5.4 inches in 2 years. We must also recall that uplifts of many feet at times of earthquakes have been recorded from Alaska, New Zealand, Assam, Rajputana, Turkestan, and Italy. Surely in some or all of these places orogenic activity is as active as in Japan and California (where our level nets happen to be more accurate), and they are as prophetic of further mountain uplift.

Of course broad warping is also going on now; perhaps as good an example as any is that of France, for this cannot be directly due to glacial loading or unloading. Schmidt (1922) has shown from precise levels that between 1860 and 1890 France has

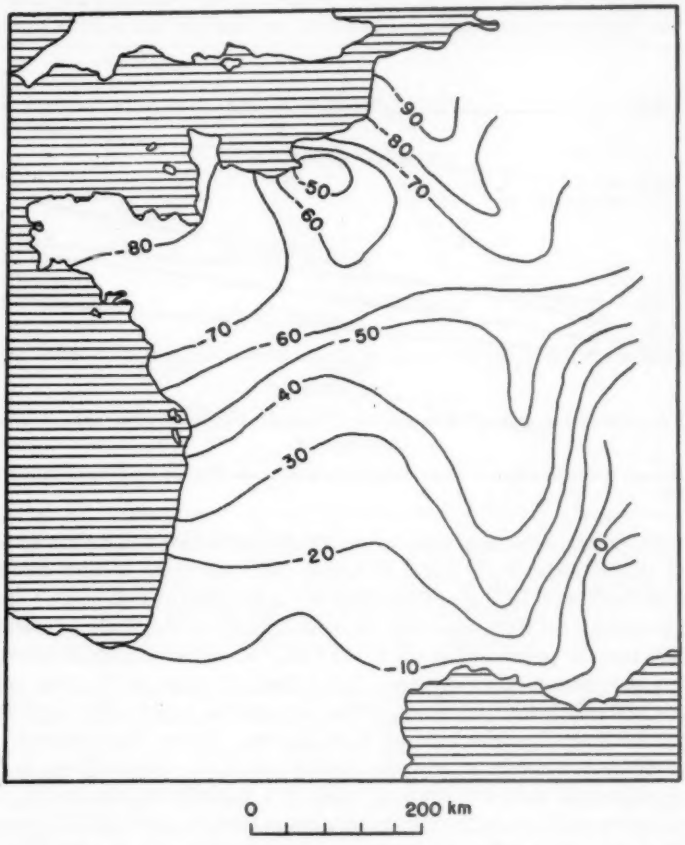

FIGURE 4.—*Map showing level changes (in cms) in France, deduced from precise levels of 1857–1864 and 1884–1893*
After Schmidt.

tilted broadly northward (Fig. 4). From the zero isobase near Marseilles the sinking steadily increased to the Belgian coast, where it amounted to about a meter. This is at the rate of about 10 feet per century—even faster than most of the localized orogenic movements we have considered. Because it involves a large area and is downwarping everyone will call this an epeirogenic movement. If continued it will bring about the epicontinental flooding of much of France.

It is true that Schmidt's interpretation of the French leveling data has been

questioned and the reality of the downwarping he inferred may be doubted. (Gutenberg, personal communication). However, this may be, the studies of modern fluctuations in sea level at various ports indicate real movements quite analogous with those inferred by Schmidt. (Gutenberg, 1941). Whether or not there is a real objective distinction between orogenic and epeirogenic movements (which I am inclined to doubt), the movements I have mentioned from California and Japan are orogenic: they are localized along definite axes, they involve stratified rocks, and in some the dips are high enough to be read with a Brunton.

Thirty years, or even a century, furnishes an absurdly short base line from which to triangulate into the geologic future. Nevertheless precise surveying should carry more weight than any amount of theory as to actual conditions. It seems to me the burden of proof is on the proponent of the idea that we are in an anorogenic period or even in the dying phase of an orogenic one. We are more justified in asserting that, although most of the crust is moving only slightly or not at all, localized belts are actively orogenic. That the present is a random sample of the geologic past and that this has always been the situation seems to me impossible to prove, however likely it may be. But the converse dogma, that the present is not a fair sample of the past, and that mountain-making epochs differ drastically from the present, seems to me quite without compelling support. Let us turn to the record of the Tertiary rocks.

THE TERTIARY RECORD IN CALIFORNIA

In Bulletin 118 of the California Division of Mines, scores of authors have described the oil fields of the State. On Figure 5 I have plotted the ages assigned by authors of 60 descriptive articles to the unconformities they have recognized in their respective fields. I have not included all the fields nor have I consulted other areal descriptions, though it is certain that if I had I would have found record of still other unconformities. The definite marks are the ages assigned by authors to the immediately overlying beds at each unconformity. The chart exaggerates the length of post-Pleistocene time, but not enough to show the three different deformations that can be recognized at one place or another around the State.

Of course this chart includes many arbitrary things. Some of the age determinations are very precise in the local time scale, with close faunal zoning of the beds above and below. Others are far less closely defined. Some authors admit dating a particular unconformity by analogy with another, either near by or elsewhere in the State. The age assignments are therefore not all independent, but are weighted in favor of certain times as those of "known deformation." That this is hazardous is shown by the fact that the Wilmington anticline, whose major uplift was Mid-Pliocene, showed practically no rejuvenation at either of the times of marked uplift of the Palos Verdes Hills, only 4 or 5 miles to the west, nor at the time of the Mid-Pleistocene folding at Signal Hill 2 miles to the northeast.

The chart nevertheless shows unconformities of 42 different ages. Now, according to Stille (1936), only six of these really count. Here he was following Ralph Reed, certainly one of the greatest authorities on California geology. Both these students

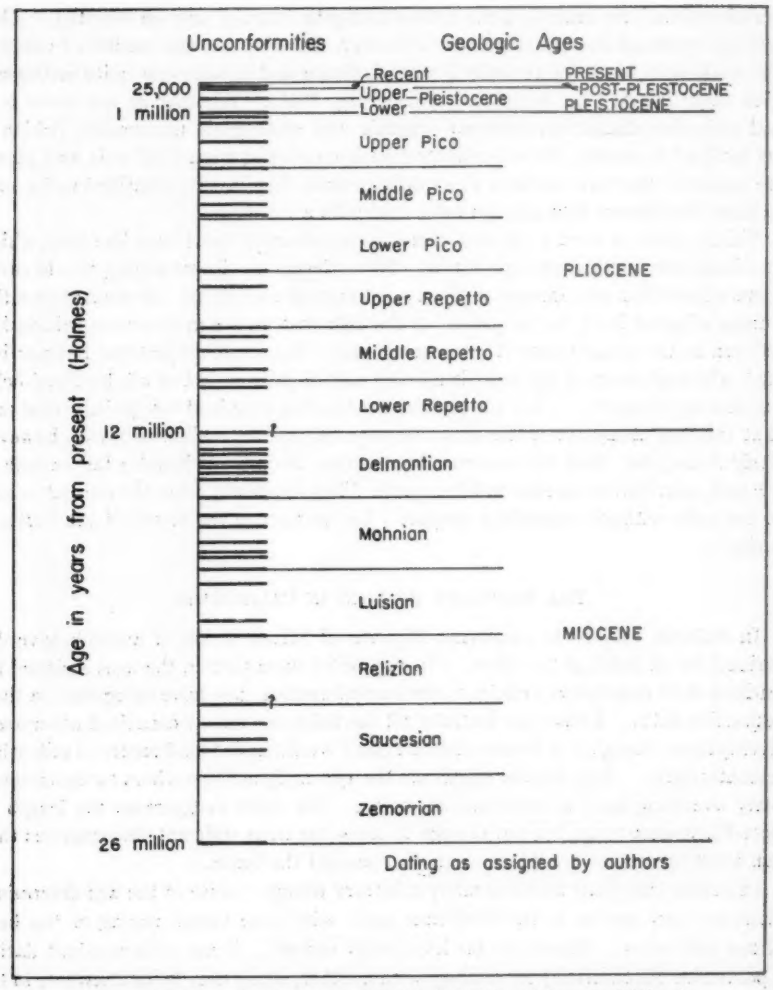

FIGURE 5.—*Unconformities reported in 60 oil-field descriptions in California*

The marks indicate age assignments of immediately overlying beds at the unconformity. Data from Bulletin 118, California Division of Mines, 1943.

took the position that an unconformity recorded from only one or two places was minor; the really significant ones are those widely represented around the State.

But, as both Von Bubnoff (1931) and Woodring (1938) have pointed out, this assumes what it sets out to test. If the movements were, in fact, going on continuously or sporadically throughout the Tertiary, obviously the places to seek evidence of discordance in strata are at the borders of the basins, not in the middle

of them. In the middle of the basins only the movements that finally stopped deposition could be recorded, and of course such movements would be most widely recorded. Movements while the basin was still drowned would leave structural records only at the borders—the places most vulnerable to erosion and loss of record. For example, it is most unlikely that there is an unconformity among the Quaternary sediments of the Santa Barbara Channel to record the mid-Pleistocene "Pasadenan orogeny" of Stille which has so pronounced a record on the Channel shore at Ventura just to the north. If we confine our attention to the middle of the basins, the only unconformity we *can* detect is the record of ultimate emergence. The movement this records might be smaller than many movements whose records are only locally preserved and thus less conspicuous, though quite as great in magnitude.

To illustrate this, let us turn to the Kettleman Hills area. Figure 6 shows the geology of the area near Coalinga and Kettleman Hills, as portrayed on the Geologic Map of California of 1938. No unconformities have been recognized in the Kettleman Hills oil field in the strata penetrated by wells, as far down as the middle Eocene Avenal sandstone, although the whole Cenozoic is represented. Yet less than 20 miles west and northwest of Kettleman Hills no less than six unconformities can be recognized in this interval. Nor was the uplift of the Kettleman Hills anticline greater than the uplifts recorded by these older unconformities in near-by areas. The structural relief of the Kettleman Hills fold is about 6000 feet; at one place or another each of these older unconformities near by may be shown to have had a comparable or greater relief. Phenomena exactly comparable are well known from both the Ventura and Los Angeles basins.

IMPLICATIONS OF THE SEDIMENTS

The implications of grade size of sediments as to the uplift of the source areas has, of course, often been noted. But it seems to me that the bearing of this kind of evidence on the timing of orogenies has not been sufficiently stressed. The structural history of the dated sedimentary rocks—in other words, of the depositional basins—has been emphasized at the cost of slighting the story that the sediments themselves have to tell about their source areas—the only clue we will ever get of structural history in areas of long-continued erosion.

For example, let us take the land-laid Sespe formation of the Santa Ynez Range, California. The Sespe has yielded vertebrate faunas of ages ranging from late Eocene through Oligocene to Early Miocene. Westward it wedges out into marine beds of Oligocene age (Bailey, 1947). It therefore represents a long time span—perhaps 12 or 15 million years. Of course no one would contend that this continental formation represents uninterrupted sedimentation—there must be innumerable diastems, even though its thickness is as much as 7500 feet. Both at top and bottom it is coarsely conglomeratic though in the middle there are many mudstones, and the conglomerates are finer and more discontinuous.

I wish merely to reiterate the truism that coarse gravel is not transported by sluggish streams. The gravel furnished to the Sespe basin at intervals through a span of 12 million years implies steep stream gradients through a time far too long for a single erosion cycle, if we are to judge by the physiographic history of the Ventura region.

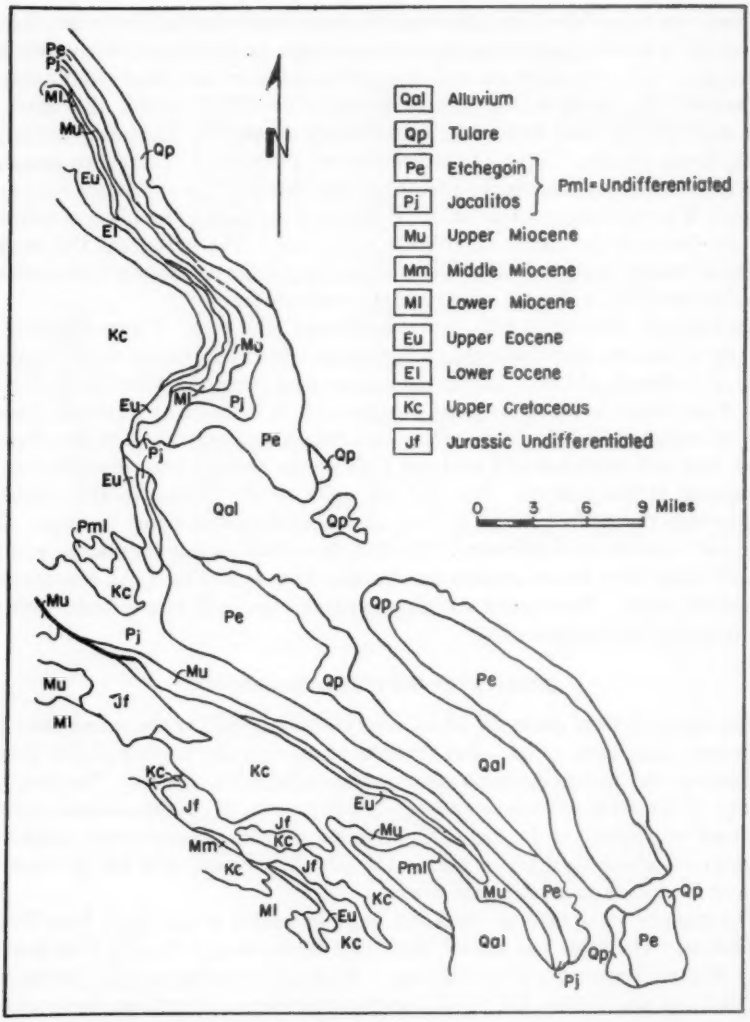

FIGURE 6.—*Geologic map of the area of Coalinga, Kettleman Hills, and McClure Valley as shown on the Geologic Map of California, 1938*

(California Division of Mines.)

The Ventura River crosses the Ventura Avenue anticline, which has been uplifted many thousands of feet since the middle of the Pleistocene. Putnam's (1942) study of the physiography showed that this uplift was episodic, it is true, but that many-times-repeated movements were involved during this short time. A thickness of more than a mile of rocks has been removed from the crest of the fold in perhaps a

million years—if we follow Stille's estimate (1936), in less than 250,000 years! Now the Ventura River gravels are coarse, perhaps about as coarse as those of the Sespe. But the several square miles of terrace in the Ojai Valley, upstream from the fold, suggest that the country would long since have been reduced to gradients far too low to supply such gravels had it not been for recurrent uplift. It seems sure that quiescence for 12 million years would have base-leveled the narrow coastal mountains, which paleogeography shows were the main sources of the Sespe sediments. The moral seems inescapable: the conformable series of Sespe conglomerates records repeated or continual uplift of the basin borders, even though no structural evidence of folding during its deposition has been found. Now, it is true that some might contend that these uplifts were epeirogenic as they have no record in unconformities. But we must recall that the source area was almost surely narrow and was uplifted more than a mile in order to retain the gradient recorded by the gravels. That such a differential uplift could take place without folding or faulting of the kind that most of us regard as orogenic seems impossible.

So much for the Tertiary of California. This is supposed to be an exceptional area and time. But I believe it is no more exceptional today than were other areas at other times in the geologic past. Let us consider next the implications of the Cretaceous sediments of the Rocky Mountain geosyncline.

Reeside (1944) has published isopach maps that allow us to compute the volume of Cretaceous sediments of the Great Plains and Rockies. We can also make reasonable estimates of the relative proportions derived from east and west, respectively. The volume of Lower Cretaceous sediments of western provenance turns out to be about 140,000 cubic miles, or enough to cover the whole United States to a depth of 250 feet. We can also make a reasonable estimate of the source area. My assumptions are shown on Figure 7, where I have taken Reeside's zero isobase as the eastern limit of the source area and have arbitrarily drawn its western limit along the western border of Idaho, continued it southward to about the latitude of Austin, Nevada, and thence south-southeastward to the Gulf of California. Such a line roughly bisects the interval between the Cretaceous of the Rockies and the even thicker sections of the Pacific Coast. The area included between these lines from the 49th parallel to the Mexican border is about 310,000 square miles. If we assume that a cubic mile of Cretaceous sediment represents denudation of a cubic mile from the source area, we find that roughly half a mile of rock was eroded during the Lower Cretaceous. The Lower Cretaceous rocks are chiefly clastic, except in the extreme south, so this arbitrary assumption seems not unreasonable. When we remember that this is an *average* figure for denudation, and we know, for example, that very little was eroded from large areas of Utah, where Upper Jurassic is widely preserved today, it is clear that other areas must have been deeply eroded. That is, there must have been high relief in large parts of the source area during the Lower Cretaceous. The sediments give little sign of secular changes in grain size and hence of planation of the source area during the roughly 25 million years that Holmes assigns to the Lower Cretaceous. To me this suggests repeated or continuous uplift in the source area.

Whatever the significance of the Lower Cretaceous may be, the comparable figures for the Upper Cretaceous allow little ambiguity. According to Reeside's isopachs the

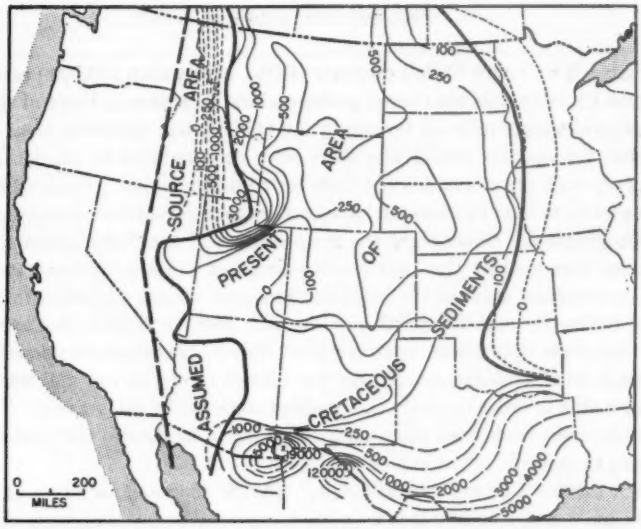

FIGURE 7.—*Thickness of Lower Cretaceous sediments of the Rocky Mountain geosyncline*

After Reeside (1944). The assumed source area is limited on the east by Reeside's zero isopach and on the west by the ine roughly bisecting the interval between the Cretaceous of the Rocky Mountains and that of the Pacific littoral.

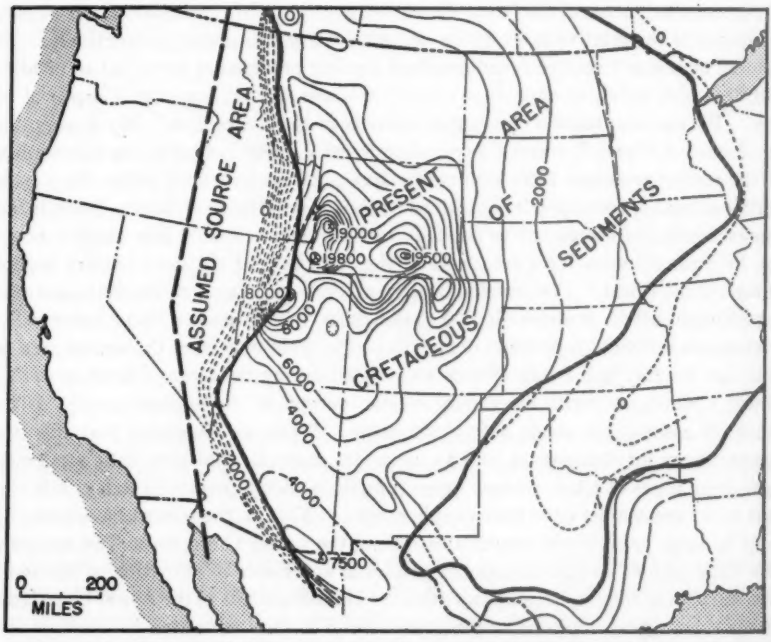

FIGURE 8.—*Thickness of Upper Cretaceous sediments of the Rocky Mountain geosyncline*

After Reeside (1944). The assumed source area is bounded on the east by Reeside's zero isopach, on the west by the line roughly bisecting the interval between the Cretaceous of the Rocky Mountains and that of the Pacific littoral.

total volume of sediment was more than 4½ billion square mile feet, or seven-eighths of a million cubic miles—enough to cover the whole United States to a depth of 1500 feet. My best estimate from Reeside's maps, from the stratigraphy, and from discussions in the literature is that nearly 90 per cent came from the west. Again using Reeside's zero isopach as the eastern limit of the source area, the same western boundary, and the same assumption of volume for volume transfer, we find that the average denudation of the source area was 26,125 feet or 4.95 miles (Fig. 8). When we remember that the source area, at least in Nevada, was largely limestone, we would be tempted to say that erosion must have been even deeper, for there is little limestone in the Cretaceous sediments, and most of it must have by-passed the preserved parts of the geosyncline. On the other hand there is an unknown, but probably considerable, volume of volcanic debris that probably does not represent denudation products, though of this we can hardly be sure because Permian and possibly early Mesozoic lavas also occur in the source area. Let us generously attribute 40 per cent of the sediments to new volcanics. We can afford to be generous—even profligate, as I feel sure this estimate would be—for we still have about 3 miles of average denudation of the source area.

Obviously this is only a rough guess. Some sediment may have come south from Canada or north from Mexico, and perhaps the western limit of the source area may be changed—though not too much or we will have difficulties with the Upper Cretaceous of California, Oregon, and British Columbia. Such corrections, if indeed they be corrections, might quite as well be balanced. Errors introduced by counting sediments twice, for some were undoubtedly reworked from older parts of the Upper Cretaceous during later parts of the period, do not appear to be significant. If they occur they are not reflected in Reeside's maps. His total for all Upper Cretaceous is larger by about 10 per cent than the sum of the volumes of the several component time zones. It is his lesser total that I have used. Whatever such modifications might be, the volume of sediment is so great and the source area relatively so small—less than 160,000 square miles—that there seems to be no escaping an average denudation approaching 3 miles.

Again we must remember that this is an *average* figure. Some areas may have been more deeply eroded and others less. To me this spells mountainous terrane.

Spieker (1946) has given us evidence for at least two intra-Cretaceous orogenic movements, one in Colorado time, the other in middle Montana time, in central Utah. He emphasizes that both unconformities fade out eastward within a very few miles into perfectly conformable sections. Furthermore, in these conformable sections there is no obvious coarsening of sediment to correspond with the great conglomerate zones to the west. All these features are, of course, precisely like those of the California Tertiary.

Now let us recall the temporal stability of the sedimentary facies in the Cretaceous of Utah and Wyoming. It is practically all clastic. After the advance of the Dakota sea the marine shales interfingered westward with marine sandstones and with continental coal-bearing beds. The transition zones from marine to continental and from marine sand to marine clay oscillated across a wide territory, shifting generally more and more eastward with time. But the lithology as a whole is remarkably

constant through time—the facies changes are geographical but taken as a whole not secular. In other words, they do not suggest peneplanation of the source area. The Cretaceous coals of the western border are of Colorado age; farther east they are of early Montana and finally of late Montana age.

Bearing in mind both this stratigraphic uniformity and the great denudation needed to supply the sediments, is it reasonable to assume that the episodes of folding and erosion demonstrated by Spieker were the only ones that transpired during the deposition of the three-quarters of a billion cubic miles of sediment of the Upper Cretaceous? Away from the actual folded area, the stratigraphic column both above and below the strata corresponding to the particular times of orogeny is essentially the same. Is it not more reasonable to assume that throughout the period sedimentation was conditioned by similar orogenic movements? If erosion had removed a belt 20 miles wide in Spieker's area, or if, in this belt, the younger sediments had been only a little less deeply eroded, his critical observations could not have been made. The fact that no other structural discordances have yet been found and dated paleontologically surely is no evidence that these were the only times of such disturbances. The great volume and the homogeneity of the sediments to the east, and their rather modest source area, make it seem probable to me that scores of similar disturbances took place during the 40 million years of Upper Cretaceous time. Our lack of direct knowledge of them is due not only to the large areas still unsurveyed but to inherent imperfections of the geologic record. To some of these we will now turn.

IMPERFECTIONS OF THE GEOLOGIC RECORD

Barrell (1917), Holmes (1947), Schuchert (1931), Stille (1936), and Umbgrove (1947), among others, have pointed out that the maximum known thicknesses of the younger series and systems are greater in proportion to their duration in years than are those of the older systems. From this they have concluded that orogeny and concurrent sedimentation have speeded up with the passage of time.

There is indeed a remarkable apparent increase in thickness of the younger series as compared with the older (Fig. 9). But before concluding from this that the rate of sedimentation has speeded up with time, I would like to consider an alternative interpretation of the data—that we simply have a more complete record of the younger strata than of the older—an interpretation that I must confess seems to me so likely that it should not be lightly dismissed. But since it has been dismissed by these able geologists, I wish to present some grounds less subjective than the *a priori*.

One factor in this problem is geological perspective. The five recognized subdivisions of the Cenozoic used in this analysis are only epochs. They average less than 12 million years in length, according to Holmes' data; the six periods of the Paleozoic average 55 million years. It is possible, of course, that this difference in average length of unit recognized expresses a tectonic history that has really been more active in later than in earlier geologic time. But in part it must surely be a function of the completeness of the record. This, in turn, is a function of the area over which the rocks of a particular epoch or period are exposed.

D. V. Higgs, a student in the University of California, has become interested in

this problem and has painstakingly determined the area of exposure of the different series and systems recognized by the U. S. Geological Survey in its Geologic Map of the United States. He has generously made his data available to me (Fig. 10).

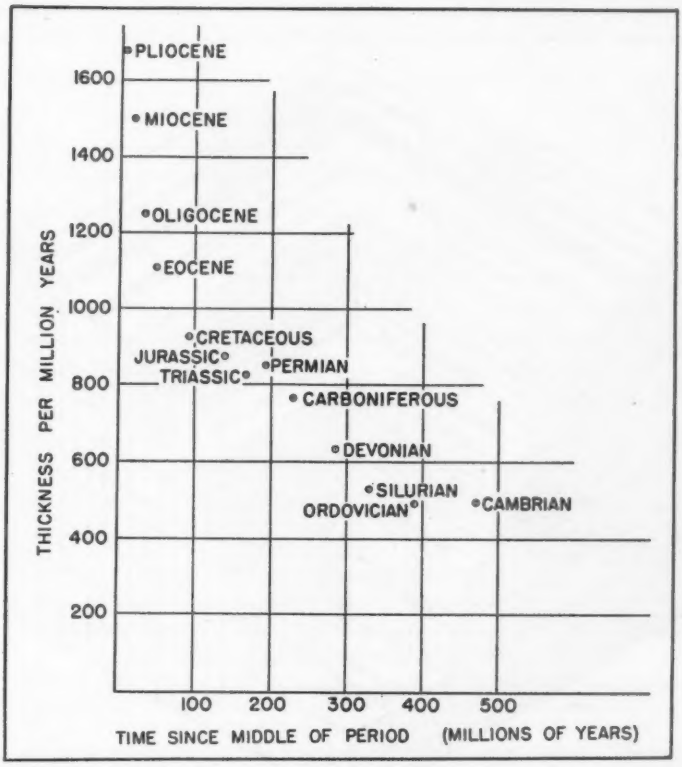

FIGURE 9.—*Maximum thickness of systems and series per million years duration of the corresponding periods and epochs plotted against their median ages*

Data from Holmes (1947).

When we recall that the Quaternary rocks cover 15 per cent of the country and mask all the older ones in their area of outcrop, the Pliocene mask all older than themselves, and so on, this block diagram suggests immediately that the chance of our seeing the record of a particular geologic phenomenon diminishes markedly with the time elapsed since it occurred. This is shown in another way in Figure 11, where the outcrop area per million years is plotted against the median age of the rocks concerned.

Clearly, the rocks deposited in a particular time interval tend to be more widely exposed than older rocks deposited in an equal time interval. Since both area and thickness are parameters of volume, this relation may be thought to support Barrell's

contention that rates of deposition have increased with the progress of geologic time. There are three reasons why this does not appear to me to be valid.

First, the thickest sediments are highly localized, and the chances of finding

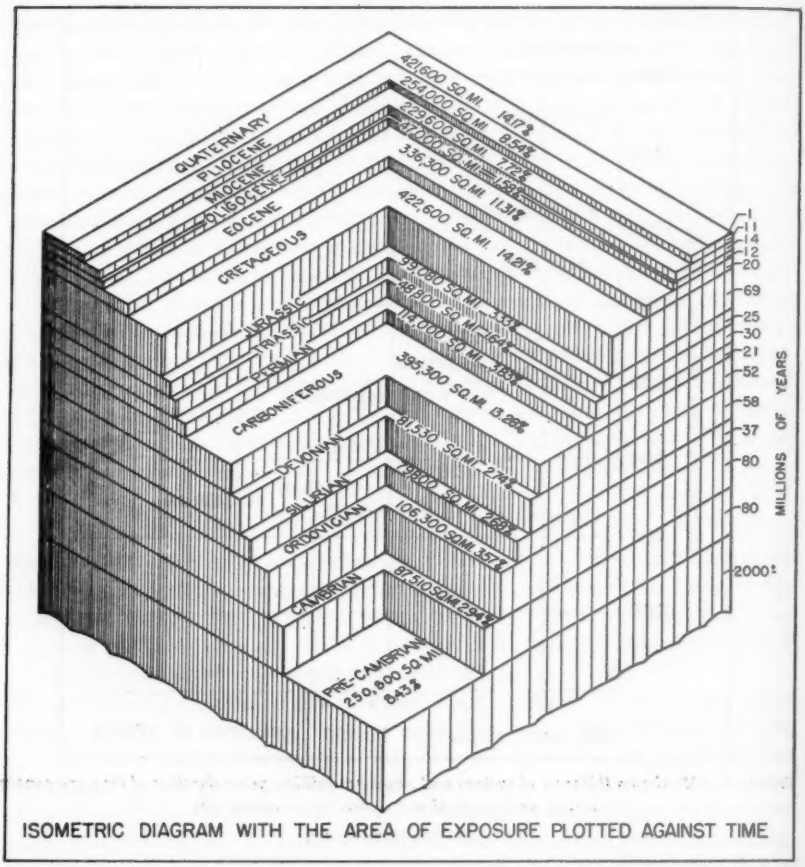

FIGURE 10.—*Block diagram showing relative areas of outcrop of the several systems and series indicated*

The area of the "tread" on top of each unit is proportional to the outcrop area in the United States; the height of the "riser" is proportional to the duration of the period as given by Holmes (1947). The outcrop area per million years duration of the corresponding time unit tends to be greater the younger the rocks involved. Courtesy of D. V. Higgs.

precisely the thickest section must diminish with age both because of burial by younger formations and because of erosion. For example, the maximum thickness of Pleistocene rocks yet recognized in a single section in the United States is that in the Ventura Basin—about 6000 feet. Even this does not represent all of Pleistocene time, for deposition ceased with a mid- or late-Pleistocene folding. Yet in comparison with the whole area of Pleistocene deposits of the United States, or even

with the marine Pleistocene, this is a trivial exposure. Over most of the area the thickness of Pleistocene would probably be a few scores or at most a few hundred feet. If some 15 per cent of the country were covered by the deposits of some "Quintary" epoch, what are the chances that erosion would expose precisely this Ventura

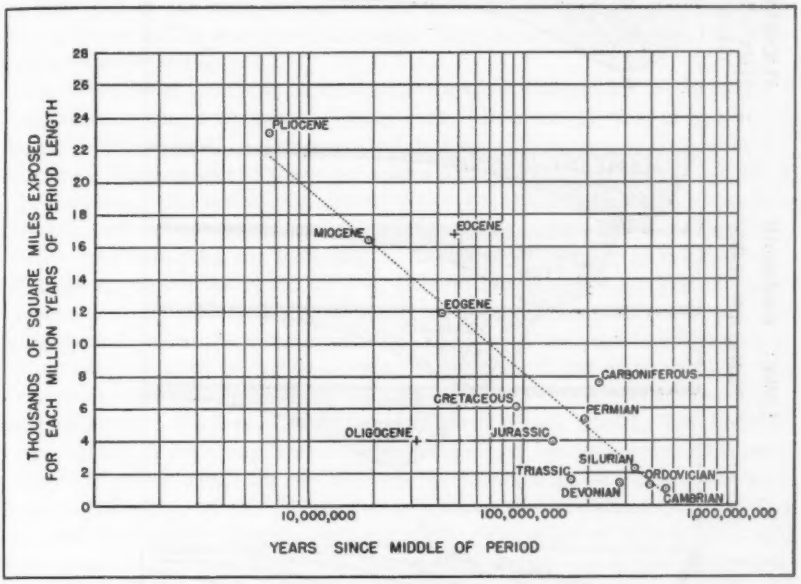

FIGURE 11.—*Graph showing area of outcrop per million years duration of the corresponding time unit plotted against the median age of the rocks*

Ages from Holmes (1947). The dotted line can be given no quantitative significance but suggests strongly that, in general, the more recently a geologic event occurred, the more likely is its record to be observed.

section—a mere pin-point in the expanse of Quaternary rocks? Similarly, the Ventura Pliocene is 16,000 feet thick, approaching the world maximum of 18,000 feet given by Holmes. Again the average is very much less. In short, the areas of maximum thickness are highly localized, just as the mountains are localized. When we consider that greater and greater proportions of the older units are either concealed by younger rocks or have been eroded to furnish them, it seems probable that we have not yet found the very thickest section of Pliocene, it is still less likely that we have found the very thickest section of Miocene, and so on, into the geologic past. The 15,000 feet of Spieker's Indianola group in Utah represents only a part of Colorado time; a few miles away the equivalent rocks are less than a third as thick. Is this 15,000-foot section the very greatest of this time interval? And where are the maximum sections corresponding to the rest of Colorado time—the lateral equivalents of which correspond to the lateral equivalents of the Indianola in grain size?

A second reason for questioning the reality of the ostensible increase in depositional rate is the effect of lumping the deposits corresponding to longer time units in the

Paleozoic compared to the shorter units of the Cenozoic. The effect is clearly to lessen the ostensible rate of deposition for the longer unit. For example, Russell (1940) has shown that at least five formations may be recognized in the Quaternary

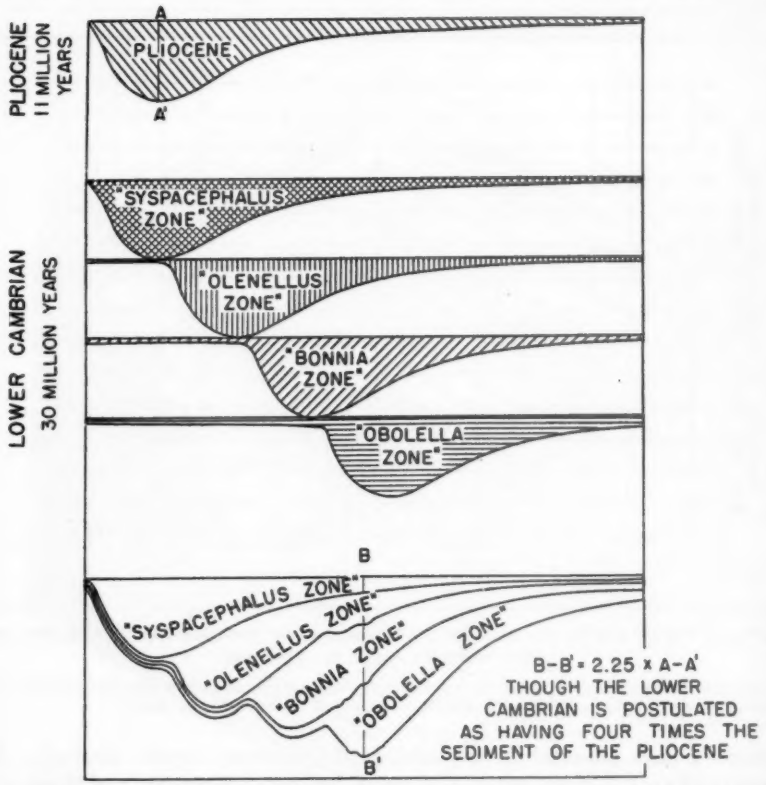

FIGURE 12.—*Hypothetical diagram to illustrate the effect of long time spans in reducing the apparent rate of sedimentation deduced from maximal sections, because of shifting loci of maximum deposition*

It is postulated that the Lower Cambrian is about 3 times as long as the Pliocene, that each of the four zones has an equal volume of sediment deposited over an equal area. Owing to the shift in site of maximum sedimentation with time, the apparent rate of sedimentation as deduced from a measured section B-B′ would be only a little more than half that of the Pliocene. It seems evident that comparable time units must be used if we are to make valid deductions about rates of sedimentation at different times in the past.

of Louisiana and that at least three of these attain local thicknesses of 3000 feet. If we add the maxima of all five formations we would have a total for the Pleistocene and Recent of 10,000 feet or more. But, even in the short time span of the Quaternary the basins of sedimentation shifted. No two of the maxima coincide. The thickness of Quaternary encountered in a well here might thus include the maximum thickness of one of the units but a much condensed section of each of the others.

Judging from Russell's structure contours, the total might be about 5000 feet, about half the sum of all the maxima. Now it seems clear to me that the maximal sections compiled by Schuchert, Goodchild, and all the others who have studied sedimentation rates through geologic time are internally inconsistent because of this factor of shifting loci of maximum sedimentation in time. By lumping together deposits of a very long time span on the same basis as those of a short time span we are consistently lowering the ostensible rate of sedimentation. I have attempted to show the effect diagrammatically in Figure 12.

Thus, if the Lower Cambrian is one third of the whole Cambrian it would be about 27 million years long, according to Holmes' scale. It is thus 2½ times as long as the Pliocene. Holmes (1947, p. 148) gives 1640 feet per million years as the sedimentation rate for Pliocene time, 500 feet per million years for Cambrian time. I think it is apparent from what we know of shifting sites in the Quaternary that this apparent difference need not correspond to a real difference of anywhere near this magnitude.

Still a third reason for not accepting the ostensible speeding up of sedimentation with time has been pointed out by Von Bubnoff (1931). The marginal zones of depositional basins are not only the sites of the thickest sedimentation, but they are also, in general, the most active tectonically. In many places unconformities present in these marginal zones die out basinward. This has been proved by mapping and still more by drilling in rocks of all ages and in all parts of the world. One of the best examples is in the lower Carboniferous of Upper Silesia, cited by Von Bubnoff. In consequence, in the young formations we see the thick, near-shore clastic deposits; the finer offshore sediments are still largely beneath the sea, as are the Santa Barbara Channel equivalents of the Pleistocene conglomerates at Ventura. Among the older rocks precisely the opposite tendency exists. The record of the coarser clastic border zones is much more liable to be destroyed by erosion than is that of the basin center. Statistically we should expect that the older the period, the more likely it is to be represented by offshore deposits.

Now this is exactly what we find. Schuchert (1931) discusses this matter at length. In his compilation of maximum sections the Tertiary is wholly clastic whereas the Mesozoic is 6 per cent limestone and the Paleozoic 42 per cent limestone. Now all students seem to agree that limestone accumulated more slowly, on the average, than either shale or sandstone. Estimates of relative speeds of accumulation diverge widely, however, from 1 to 5 to 1 to 25. Schuchert summarized the matter very well when he said that as yet we have no quantitative data sufficient for estimating either rates of sedimentation or of denudation for the earth as a whole. But it is clear that a limestone section cannot represent the maximum deposit of its time span.

Holmes (1913), Leith and Mead (1915), and F. W. Clarke (1910) have all computed the ratios of clastic to carbonate rocks, assuming that the sediments are all derived ultimately from the average igneous rock. The average of the three estimates is: shale, 75 per cent; sandstone, 13 per cent; and limestone, 8 per cent. From descriptions of stratigraphic sections totaling about 130 miles, Leith and Mead estimated that the actual distribution of sedimentary rocks was: shale 47½ per cent, sandstone 32½ per cent, and limestone 20 per cent. Schuchert assumed ratios of shale 50 per

cent, sandstone 32 per cent, and limestone 18 per cent. He therefore believed that the limestone thicknesses should be multiplied by 4 to get the true maximum thickness deposited during the corresponding time interval. He tested this with the Caribbean Cenozoic rocks. By multiplying the thickness of carbonate rocks by 4 he gets a total of 54,000 feet of hypothetical sediments in the Caribbean to compare

TABLE 1.—*Maximum sections now known, converted to hypothetical thicknesses if carbonates are converted to equivalent clastics*

1	2	3	4	5	6	
Era	Maximum sediments measured (feet)	Limestone included in the maximum sections	Clastics plus 4 times the carbonate thickness	Clastics plus 6 times the carbonate thickness	Theoretical maximum assuming constant rate of deposition and that real maximum of Cenozoic is 61,000 feet	Deviation of "corrected thickness" (Col. 5) from that expected on assuming uniformity (Col. 6)
Cenozoic (58 million years)	61,000	0	61,000	61,000	61,000	—
Mesozoic (124 million years)	86,600	5,000	101,600	111,600	130,000	14%
Paleozoic (324 million years)	111,000	47,500	253,500	348,500	345,000	1%

with the 61,000 feet of clastics measured in California. To apply Schuchert's method to the Paleozoic sections seems to me quite justifiable. But since his factor of 1 limestone equals 4 clastic does not work out for the Cenozoic, I have chosen Holmes' ratio of 1 limestone equals 6 clastic. Table 1 illustrates the results of this "correction".

Now it is obvious that with all these arbitrary assumptions Table 1 (which is graphed in Fig. 13) is not to be taken as more than a suggestion of the true thicknesses of the several systems and series. However, I submit that it does indicate that there is nothing in the basic data on thicknesses to support the contention that sedimentation has been speeding up with time. When we take account—as we must—of the differences in lithology I see no reason for regarding the Cenozoic as anything but a perfectly normal era, differing not at all from the earlier ones in rates of sedimentation. The five points on the radioactive scale that Holmes regards as best established are not so many nor so well dated paleontologically as to force us to accept the durations of the individual periods as accurate within 10 per cent. Nor is it reasonable to suppose that the thickest sections have yet been found. All in all, there seems to me no valid reason for suspecting that sedimentation was any slower in the Paleozoic than in the Cenozoic. For all we know, it may have been even faster.

Nor is there any reason to doubt the validity of some sort of correction to the measured thicknesses because of the limestone contained in them. Surely the limestone deposited in any particular period must have been removed from the land surface during that time. In the Cambrian the sea must have been essentially saturated with calcium carbonate at shallow depths, just as it is today. Every foot

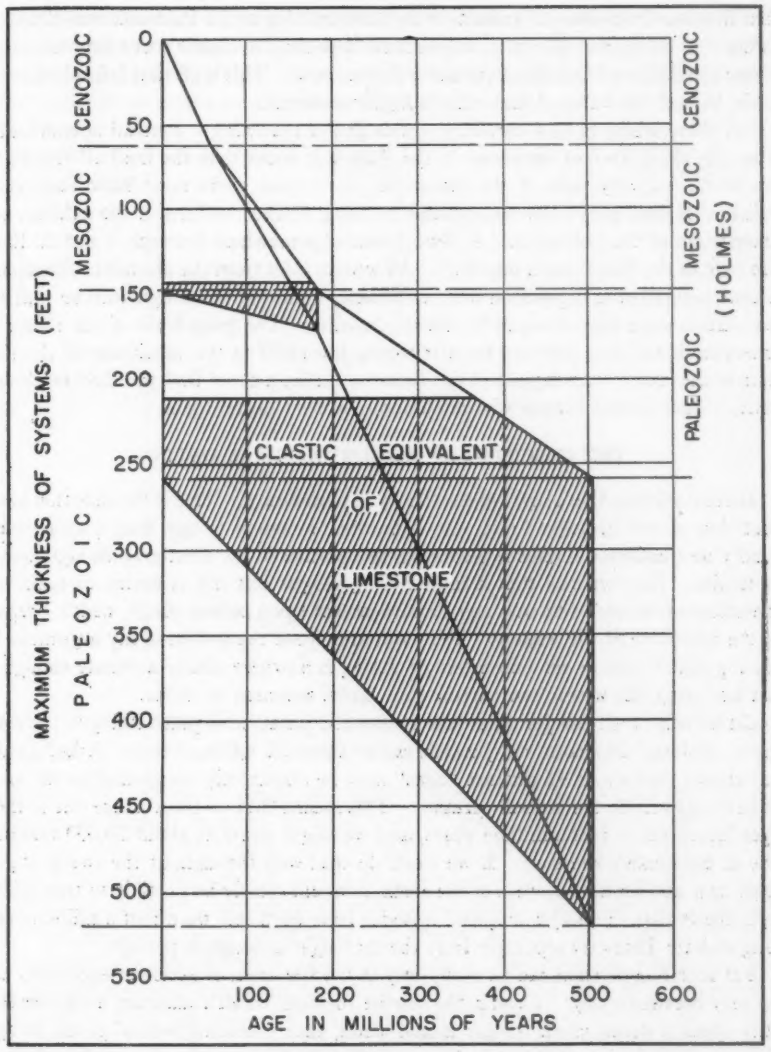

FIGURE 13.—*Graph showing effect of correcting maximum measured sections for their content of limestone*

Upper line from Holmes (1947), nearly straight line shows maximum thickness as it would be if the limestone included in the measured sections were replaced by 6 times its thickness (its rough equivalent in clastic rocks). Limestone in measured sections shown at left (from Schuchert); expanded to right by factor of 6.

of limestone deposited during a given time must represent a good many feet of denudation from an equivalent area, in order to supply the calcium equivalent to the sea. There is no reason I can think of or find suggested in the literature for thinking

that limestone was more abundant on the land draining to the Paleozoic seas than it is today. If it was not, the ratio of limestone deposited to clastic rocks deposited elsewhere should have been about the same then as now. This is all that is implied in the table, though the value of that ratio is highly uncertain.

But there seems to be a corollary of this line of reasoning. Instead of concluding from the abundance of limestone in the Paleozoic rocks that the land all stood at a low level, the very bulk of the limestone proves that there must have been great volumes of land above the sea, capable of being leached to furnish the calcium and magnesium of the sediments. A denudation of somewhere between 5 and 20 times the bulk of the limestone is required. All we can infer from the abundant limestones is that their sites of deposition were sufficiently removed from highlands so that the carbonates were not swamped by clastic deposits. The great bulk of the Paleozoic limestone is no more a reason for attributing low relief to the continents of the time than is the recent lime deposit of the Bahama Banks a proof that the Rockies do not exist, or that Mount Mitchell is only a bump on the Coastal Plain.

THE PRECISION OF PALEONTOLOGICAL DATING

There is still another aspect of the theory of periodicity. This is the question of the precision of our geologic dates. It is hardly too much to say that geology could hardly be considered a science before the recognition of the stratigraphic significance of fossils. The remarks that a mere field geologist has the temerity to make are therefore not intended to cast any disparagement upon paleontology, which I regard as the backbone of the science. Yet I will anticipate the details of my argument by stating that I believe our paleontologic dating is nowhere nearly accurate enough to bracket orogenies within the narrow time limits assumed by Stille.

Let us start with the Jurassic—the ammonitic paradise of paleontology. According to Holmes' estimates the Jurassic lasted about 25 million years. Arkell (1933) has shown that about 45 distinct "ages" may be objectively recognized in the areas of best exposure in the British Jurassic. This means that on the average one of these ages lasted about half a million years, and we might guess at about 50,000 years for one of Buckman's hemerae. If we could do that well throughout the sweep of geologic time and for all the areas of the earth, it would clearly be possible to test Stille's hypothesis that the major orogenic episodes have been less than half a million years long and are distinctly separable from the far longer anorogenic periods.

But such fine division can be made only in the few areas of excellent exposures and by very detailed study. Even in the marine Jurassic Arkell's summary indicates that only about a dozen much thicker faunal zones, each including several of the 45 ages that can locally be distinguished, can ordinarily be recognized over the whole of Britain. Thus even in the Jurassic—the most favorable of the periods, and in Britain, one of the most thoroughly studied areas in the world—the time units that can be consistently used average roughly 2 million years in duration. I believe no paleontologist will claim that the Jurassic of America can be correlated either intracontinentally or with that of England with an accuracy even approaching this. As Muller and Ferguson (1939) point out, many of the Mesozoic zones of Nevada can be correlated with those in India with more confidence than with some of our own Rocky

Mountain formations. To a nonpaleontologist it appears that for much of the world we are fortunate to make a six-fold subdivision of the Jurassic, making the consistent divisions nearer 4 million years than 2 million, even though in favorable areas with highly fossiliferous strata we can do much better.

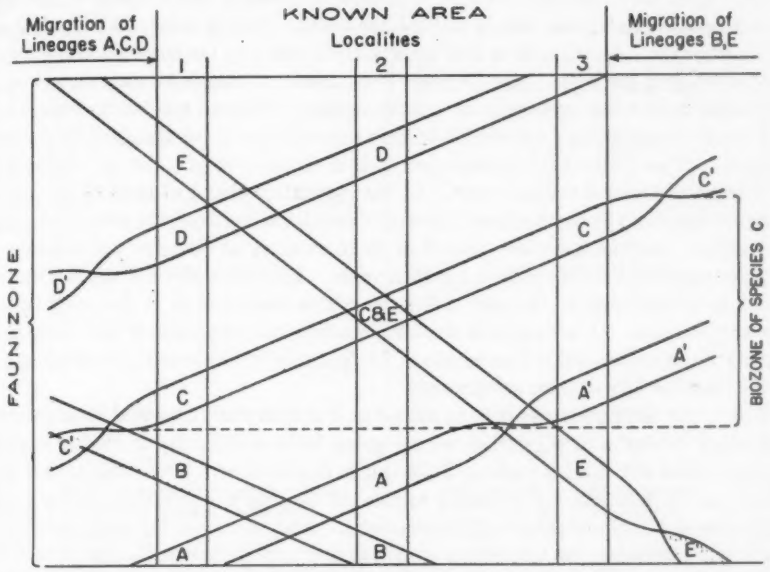

FIGURE 14.—*Diagram illustrating the relations between faunizones, biozones, and epiboles*

After Arkell (1933, Fig. 1, p. 33).

But even in fairly small regions like western Europe, where the same species may be recognized in their appropriate zones over the whole area, there are limits to the precision of stratigraphic correlation. As shown by Spath (1931) and Arkell (1933, p. 30–34), the time of migration of a particular ammonite species from its center of dispersal, though short when compared with the whole span of existence of the species, is still not negligible, even geologically. If several species that arose in different centers eventually come to live in a particular area, it is not only conceivable, but has been shown by careful collecting, that their order of superposition may be different in different parts of the area. In other words, the 50,000-year hemerae we guessed at are units far too short for distant correlations (Fig. 14). When we take into account the entire fauna we never find such reversals, but the fauna as a whole does not, of course, change so quickly as the individual species constituting it. These reversals therefore prove that even in theory it will never be possible to be certain of the precise equivalence of beds in widely separated regions. In fact, the finer the sub-divisions the paleontologist recognizes, the more likely he is to find such reversals in superposition. This point is philosophically crucial, for all our geological histories are based on paleontology, and it can therefore never offer us a chronology capable of indefinite refinement.

When we think of the many places with only poorly preserved or long-ranging fossils, of the faunas with very restricted habitats, of the large areas covered only by hasty reconnaissance, and of the still great areas that have never been really investigated at all, it is questionable whether the strata in most of the geologic column of the United States can be correlated with those of China or South Africa or Europe with a probable error less than 4 or 5 million years. This is nearly half the length of the Pliocene. An example of this uncertainty is given by the correlation chart for the Cenozoic of the West Coast (Weaver *et al.*, 1944). Though 19 stages are recognized and there is fair agreement as to local sequence, different specialists diverge by two whole stages in their correlation of the same rocks with the standard European sections. If we guess that the stages are each of the same length, this is a difference of opinion of about 6 million years. In the correlation charts of most of the other systems, drawn up by the National Research Council Committees, the paleontologists responsible recognize anywhere from 9 or 10 to as many as 30 zones and subzones. The average length is then around 4 million years. Every one of these charts is introduced by a disclaimer on the part of the committee concerned as to the precision of the correlations. As a complete outsider, reading the annotations and minority reports of the several committee members, I infer that no one has complete confidence in the immutability of these assignments.

Now if our clock measures only in units 3 or 4 million years long and its markings are rather blurred and subjective, we are going to have difficulty in timing events within a third of a million years as Stille thinks can be done. If we wish to test his theory we must confine our attention to areas of detailed study and unusually good exposures and to abundantly fossiliferous rocks. And, of course, the more restricted the area we consider, the more likely we are to find a sporadic distribution of unconformities, separated by long records of quiet sedimentation. I have already cited the situation at Kettleman Hills, where only one episode of deformation is recorded during the entire Cenozoic, yet within a radius of 20 miles six others can be demonstrated for the same time and, in a radius of 200 miles, at least 41 others. If it be granted that there are still long periods for which we have no direct evidence of orogeny, this negative evidence cannot carry much weight. As Spieker pointed out, three additions must be made to Stille's table in the Mesozoic of Nevada and Utah alone, and in passing I can cite still another deformation (Aptian) from the Little Hatchet Mountains of New Mexico as inferred from Lasky's (1947) work. In other words, with the progress of detailed mapping the number of orogenic phases is constantly increasing, and there still are far greater areas of the earth not mapped in detail than are thus mapped.

Before leaving this question, two further aspects may be mentioned. One is that every unconformity is a definite break at any particular place though it may involve a very great difference in age at different places. Now time and trangressions and regressions of the sea are always progressing, but when we collect fossils we find our friends the paleontologists nearly always referring them to a definite pigeonhole. When a series is conformable the natural tendency of stratigraphers is to refer it all to one or another of the recognized pigeonholes, especially if the conformable beds either rest on an unconformity or are cut off by one (Fig. 15). We are all familiar with

comments of paleontologists that may be paraphrased as follows: "Though the fauna of the lower part of the formation has many affinities with that of the Carboniferous, the gradual transition to the fauna of the upper parts shows that it is really Permian for characteristic Permian fossils are abundant higher in the unbroken

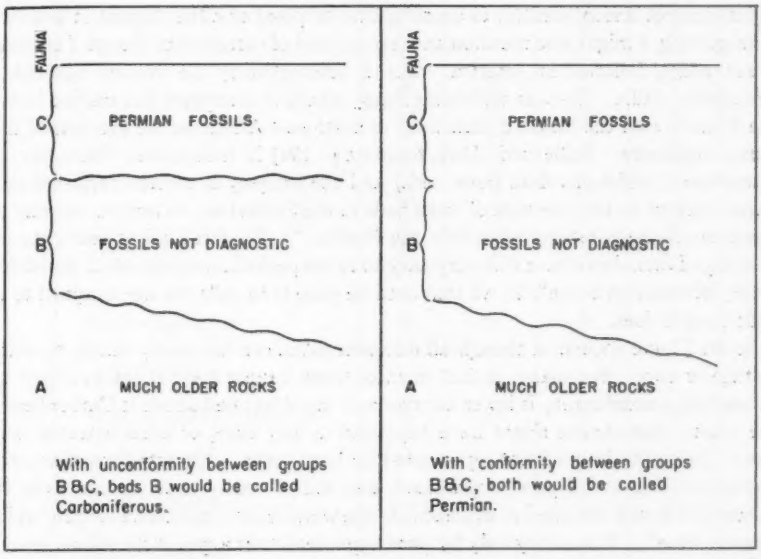

FIGURE 15.—*Hypothetical diagram illustrating probable age interpretation of two sections containing identical fossils, but one of which contains a recognizable unconformity*

succession." The beds will then appear in a stratigraphic table as Permian and date the unconformity below as pre-Permian. Now, if the fossils were identical but the series had been broken by even a minor unconformity, it is more than likely that the beds below the minor unconformity would have been called Carboniferous following their faint intimations of affinity. The lower unconformity would have been dated as late Carboniferous, and the minor unconformity would be recognized as the base of the Permian.

Now, let me hasten to say that I am not criticising this procedure. Ordinarily our correlations are based on weighing of similarities of faunas, not on identities—and the way I have just outlined is as good as the data often justify. Nevertheless, I submit that it introduces a systematic error. The dates assigned to unconformities do not *range at random on either side of the most probable age as inferred from the fossils alone* but are systematically displaced in the direction of the stage boundaries. The *named* stages thus come to appear in the written record as more distinct than any paleontologist of my acquaintance will contend that they really are. This comes from applying the excluded mean to phenomena of continuity. It is just as fallacious in geology as it is in Zeno's paradoxes. And to turn the whole wheel—identify the

unconformity from the neighboring fossils (as we must) and then cite the unconformity as evidence for rock correlation (in other words, as evidence of the ages of the same fossils)—as is implied by the proposal to use diastrophism as the fundamental basis of correlation seems to me to constitute a perfect example of circular reasoning. I have too much respect for the fossils and far too little faith in field evaluations of the significance of unconformities to consider this proposal as either logical or practical.

In passing, I might also mention another method of correlation, though I believe it is not widely followed in America. This is illustrated by the avowed advocate of periodicity, Stille. He was criticising Haug, who had correlated the marine beds of the Piraeus with the Pontian (Sahélian) of northwest Africa on the grounds of their faunal similarity. Stille said (1924, footnote p. 191) in translation, "However, the discordance (which involves these beds) and the orogeny it records furnishes additional support to the reference of these beds to the Sarmatian, so long as orogeny has nowhere else been recognized within the Pontian". By this kind of reasoning and coloring of faunal evidence it is very easy to make periodic orogeny seem plausible—if the deformation doesn't fit, all that need be done is to shift the age assigned to the beds until it does.

So far I have spoken as though all unconformities can be closely dated, to within a stage or two. But many, in fact most of them, cannot be. If the youngest bed beneath an unconformity is lower Silurian and the oldest bed above it Carboniferous, the crustal disturbance might have happened at any time, or been repeated many times during the interval not represented by local rocks. Nevertheless, many such unconformities, as Shepard emphasized, and indeed many that are not even this closely bracketed, are cited as evidence of "the world-wide Caledonian orogeny of late Silurian time". It is surely only by pure hypothesis that a gap of 60 million years in the record can be considered to prove that the deformation it records lasted less than a million years and was simultaneous with that in another area which can itself be dated only within, say, 20 million years.

Blackwelder (1909) has emphasized that, even where an unconformity can be closely dated in one locality, we cannot infer from this that it is of the same age throughout its extent. Bertrand long ago showed that uplift has been repeated many times along the same axes in northern France. Arkell has demonstrated nine uplifts of a single anticline during the Jurassic of England (Fig. 16). With which of these shall we correlate an unconformity between Portlandian and Lias in California? The thinning out of beds by erosion and their eventual overlap by a younger unconformable formation does not by any means indicate that the unconformity farther toward the area of greatest uplift was formed at the same time. Yet it is often assumed that because an unconformity can be closely bracketed in one place—the beds above and below being only a stage or so apart—the same age applies to the deformation where the younger formation rests on far older rocks. This assumption —again a natural one in absence of positive evidence to the contrary—naturally leads to the diminution of the number of disturbances to be recognized. Where the time gap is great the unconformity might actually represent half a dozen older uplifts, and only a small part of the total structural relief may have arisen within the closely bracketed limits indicated where the time gap is small. This is not simply a general

deduction; it has been proved at many places in California and elsewhere. Such proof from surface studies alone demands definite geometrical relations between the line of outcrop and the unconformity. It is therefore obvious that there must be

Periods / Names of Axes	PURBECK–WEYMOUTH	MENDIP–PEWSEY	MARKET WEIGHTON	(WILLOUGHBY–LOUTH)
PRE–APTIAN	+	+	+	+
TITHONIAN				
KIM. PORTLANDIAN	+?	−	+	+?
CORALLIAN	−	−	+	−?
OXFORDIAN	−	−	+	−?
BATHONIAN	−	+	+	−
PRE–VESULIAN	+	+	+	−
PRE–BAJOCIAN	+	+	+	−
UPPER LIAS	+	+	+	−
MIDDLE LIAS	+?	+	+	−
LOWER LIAS	?	+	+	−
RHAETIC	?	+	?	−?

FIGURE 16.—*Uplifts during the Jurassic of England*

After Arkell (1933, p. 87).

many more such composite uplifts than have so far been demonstrated. Many have been proved by well drilling, and many others must still be concealed beneath younger rocks.

SUMMARY

In conclusion, I believe our clock graduations for intercontinental correlation are only exceptionally as short as 2 million years—for most of geologic time they are more

probably 4 or 6 million years long. With such clocks it is clearly impossible to prove contemporaneity of strata—let alone deformation which cuts pieces out of the record—within a limit of half a million years or less.

It seems reasonable to take the California record at its face value, as indicating uplift at one place or another almost continually throughout the Cenozoic. Such movements were doubtless sporadic, like the modern faulting that causes earthquakes. They proceeded now quickly, now slowly, now in this area, now in that. At each locality we have definite unconformities as local conditions changed from erosional to depositional. These unconformities mark sharply defined time gaps at this particular place. But such unconformities, when traced along the strike, do not mark the identical time and indeed may change markedly in time value. The sharp record of an unconformity does not, in logic, require us to believe that it everywhere marks the same sharp instant of geologic time, and indeed it cannot. It will require more accurate dating to demonstrate the thesis of periodic diastrophism, and there are logical grounds for doubting that it can ever be done to the precision assumed by some periodicists, at least until new techniques of measuring geologic time are developed.

Finally, let us get a perspective on the basic data available to us on this problem. All the lands of the earth occupy only 29 per cent of its surface. Our geological generalizations are based upon a detailed knowledge of probably 10 per cent of this land area. For the rest, and far the greater part, we have only reconnaissance, much of it very crude. Our geologic knowledge of the far greater areas beneath the sea is so scant as to be nearly negligible. Indeed, such hints as we get from submarine topography, from marine sediments, from geophysics at sea, and from the bore holes such as the deep oil tests in the Bahamas may well make us doubt the venerable dogma of the permanence of continents. Does the submergence of Appalachia suggested by the work of Ewing and his collaborators (1937) offer greater physical difficulties than the uplift of the Colorado Plateaus? Once we grant the strong possibility that such interchanges of land and sea are possible, we cannot even be sure of the generally accepted idea that the present time is exceptional in respect to its greater-than-average emergence of the continents. For example, the present shelf seas cover nearly 5 per cent of the earth's surface. The epicontinental seas of the Upper Cretaceous flooded nearly half the present area of North America. But this is only about 2 per cent of the earth's surface. Can any geologist be sure today that this 2 per cent was not made good by emergence of lands now drowned?

Put in this frame of reference, I can see no strong reasons for believing either that diastrophism is speeding up with time or that the present time is abnormal. During our own lives we are witnessing earth movements fully vigorous enough to produce any mountains, present or past, of which we have knowledge. At the same time we are witnessing movements exactly analogous to those that brought about the epicontinental floods. Other great areas of the earth seem to be essentially stable. I know of no evidence incompatible with such a state of the crust throughout all geological history. As Spieker has shown, the epicontinental expansion of the sea in Colorado time was contemporaneous with strong orogenic movements on the very border of the sea.

Of course I do not contend that intense crustal deformation has been either con-

tinuous or ubiquitous. Doubtless there have been variations in the rates from the statistical average, and perhaps present conditions deviate from the average in one direction or the other. But it seems to me that the stratigraphic evidence indicates that these deviations have been minor—perhaps half or twice the present rates but almost surely not 10 times. The only variations demanded bythe record could have been small. World-wide orogenic revolutions do not appear to me to have been demonstrated, and I can see no grounds whatever for assuming any increased tempo for diastrophism during post-Lipalian time.

The uniformitarianism of Lyell seems not yet to require any of the amendments that have been suggested. Bullard (1944, p. 79) has said, "that sound guide, Occam's Razor, suggests that we should not wantonly assume changes [in physical laws] until they are positively required by observation or theory". I believe we should, as geologists, adopt a like attitude and not assume changes in actualistic geology until these are compelled by observations. It is my belief that such changes are not yet required by the evidence at hand. Long live Lyell and his doctrine of uniformitarianism!

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KARROO DOLERITES OF THE UNION OF SOUTH AFRICA

BY FREDERICK WALKER AND ARIE POLDERVAART

CONTENTS

	Page
Abstract	594
Scope of investigation	594
Introduction	594
Previous work	595
Course of present investigation	596
Acknowledgments	596
Field characters	598
Geological structure of the Karroo Basin	598
Distribution of Stormberg lavas and summary of their field characters	600
Drakensberg lavas	600
Zuurberg lavas	601
Lebombo lavas	601
General distribution of Karroo dolerites	602
General observations on field relations of Karroo dolerites	603
Habit of intrusions	603
Sills	603
Curved sheets	605
Dikes	606
Abnormal dikes	607
Multiple intrusions	610
Bell-jar intrusions	610
Laccolith	610
Pegmatitic schlieren and residual injections	611
Jointing and mode of weathering	612
Behavior of the dolerite magma toward associated sediment	613
Mobilization and metasomatism of sediments	614
Petrography	615
Classification	615
Chilled modifications	623
Feldspar-rich modifications	624
Order of crystallization	624
Sediments	625
Metamorphosed sediments	626
Assimilation products	627
Rheomorphic veins	628
Mineralogy	629
General statement	629
Methods	629
Nomenclature	631
Olivines	633
Feldspars	634
Orthopyroxenes	635

	Page
Pigeonites	638
Clinopyroxenes of high optic axial angle	640
Accessory minerals	646
Biotite	646
Amphibole	646
Iron ore	646
Apatite	647
Analcite	647
Chlorophaeite	647
Iddingsite and bowlingite	647
Prehnite	647
Chemical data	647
Petrology	648
Comparisons with other magmatic provinces	648
Differentiation of the Karroo magma	650
Trace elements	654
Iron enrichment in basalt magma	656
Dolerite pegmatites	662
Later stages of crystallization	664
Production of tholeiites from olivine dolerites by the incorporation of resurgent volatiles	665
Origin of the Hangnest, Hanover, and Kentani types	667
Hangnest type	667
Hanover type	668
Kentani type	669
Basaltic modifications of Karroo dolerites	671
General statement	671
Chilled contact basalts	672
Late basalt veins and dikes	674
Reaction of dolerite magma with sediments	674
General statement	674
Volatile content of Karroo dolerite magma	675
Hyperfusible components of the Karroo sediments	676
Physico-chemical aspects of the problem	677
Chemical data	679
Discussion	679
Iron-rich phase of emanations	680
Sodium-rich phase of emanations	680
Potassium-rich phase of emanations	681
Other examples	682
General observations	682
Assimilation and pure melting	684
Mechanism of intrusion	684
General statement	684
Dolerites and lavas within the Karroo Basin	684
Intrusion of the Karroo dolerites	685
Outlying areas in Southern Africa	687
Trias-Lias vulcanicity in Gondwanaland	687
Karroo igneous phase in relation to Gondwanaland	689
Weathering of Karroo dolerites	690
Macroscopic features	690
Microscopic features	692
Appendix—Drakensberg, Zuurberg, and Lebombo lavas	693
Introduction	693

Page
Petrography 693
Drakensberg lavas 693
Zuurberg lavas 693
Lebombo lavas 693
Chemical data 694
Bibliography 694

ILLUSTRATIONS

Figure Page
1. Distribution of Karroo System in the Union of South Africa 597
2. Distribution of Karroo System in Africa 600
3. Map and section of Bulthouders Bank sheet, Beaufort West 604
4. Map and section of Glen Grey sheet, Queenstown 605
5. Map and section of Insizwa sheet, Mount Ayliff 606
6. Map and section of Elephant's Head dike, Matatiele 608
7. Map and section of New Amalfi sheet, Matatiele 609
8. Map and section of Mount Arthur complex, Kokstad 611
9. Sketch of Karroo "koppie" types 613
10. Section of Hangnest sill, Calvinia 615
11. Variolitic picrites 620
12. Karroo granophyres 620
13. Iron-rich differentiates of New Amalfi (Matatiele) 621
14. Tholeiites of Mount Arthur (Kokstad) 621
15. Hornfelses of Insizwa (Mount Ayliff) 626
16. Assimilation product at Hangnest (Calvinia) 628
17. Assimilation product at Mount Arthur (Kokstad) 628
18. Optic properties and chemical composition of orthopyroxenes 630
19. Course of crystallization and γ curves of clinopyroxenes 631
20. Nomenclature of plagioclase, olivine, and orthopyroxene series 632
21. Nomenclature of clinopyroxenes 632
22. Zoning in plagioclase of Insizwa (Mount Ayliff) 634
23. Pigeonite-orthopyroxene stability relations 637
24. Relation of orthopyroxenes, pigeonites and other clinopyroxenes 638
25. Distribution of analyzed dolerites in the Union of South Africa 648
26. Chemical variation of Karroo dolerites 653
27. Differentiation trend of Karroo dolerites 657
28. Trends of differentiation of Skaergaard, Elephant's Head New Amalfi, and Insizwa intrusions 657
29. Differentiation trend of Tasmanian dolerites 658
30. Differentiation trend of Palisade diabase 658
31. Differentiation trend of Siberian traps 659
32. Differentiation trend of late Paleozoic dolerites of Britain 659
33. Differentiation trend of Hawaiian lavas 660
34. Differentiation trend of the Garabal Hill—Glen Fyne complex 660
35. Differentiation trend of the Bushveld igneous complex 661
36. Variation of dolerites and dolerite pegmatites 662
37. Sketch of xenolith at Alewyn's Gat (Beaufort West) 681
38. Section of Rietkop sill (Calvinia) 682
39. Variation of dolerites, dolerite pegmatites, granophyres, and sediments 683
40. Section of the northern Natal Coalfield 686
41. Reconstruction of Gondwanaland during Trias-Lias times 688
42. Relationship of Africa and Antarctica at the close of the Karroo period 689
43. Isohyets for the Union of South Africa 691

Plate — Facing page
1. Lava and sill escarpments — 602
2. Dike and sill features — 603
3. Jointing and weathering of Karroo intrusions — 612
4. Pegmatite schlieren and weathering phenomena in sills — 613
5. Contrasted weathering of dolerite types — 614
6. Xenoliths and rheomorphic veins — 615
7. Photomicrographs of Karroo dolerite types — 702
8. Photomicrographs of dolerite differentiates and associated rocks — 703
9. Photomicrographs of contact basalts, sediments affected by dolerite, and hypersthene with graphic intergrowth — 704
10. Metasomatic and anatectic phenomena in the Karroo — 705

ABSTRACT

The Karroo dolerites represent the intrusive phase of the early Jurassic basalts which built up the Basutoland lava plateau. Those in the Union of South Africa are described.

Both basalts and dolerites are of tholeiitic magma type and are characterized by low Fe_2O_3 and alkalies. Most of the dolerites show little differentiation, but olivine-rich types and acidic veins and schlieren are recorded. The differentiation trend was toward moderate iron enrichment. The magma was extremely active toward the associated sedimentary rocks causing widespread rheomorphic and syntectic phenomena. Many of the rocks previously described as "diorite" or "granophyre" were found to be transfused siltstones.

The dolerites occur chiefly as sills, dikes, and inclined sheets, but "bell-jar" intrusions occur near the focus of activity southwest of Basutoland.

The mechanism of intrusion is discussed. The authors believe that the Karroo dolerites were injected under conditions of tension following the buckling of the Samfrau geosyncline and represent only one phase of the pronounced igneous activity which spread over a large portion of Gondwanaland during Trias-Jura time.

SCOPE OF INVESTIGATION

INTRODUCTION

Toward the close of Triassic time the Southern Hemisphere was in a state of tension which gave rise to volcanic outpourings on a colossal scale. Vast floods of basalt built up a series of great lava fields in South America and Southern Africa, while lesser outbursts of igneous activity broke out in Australia and Tasmania, New Zealand and Antartica, as well as numerous regions in the Northern Hemisphere. According to the interpretation of Du Toit (1938, p. 94) this activity was (in the Southern Hemisphere) a result of the loosening of the bonds holding Gondwanaland together. His reconstruction of Gondwanaland is given in Figure 41, which also shows the distribution of the Trias-Lias vulcanicity.

There was an even greater penecontemporaneous hypabyssal phase during which sheets and dikes of dolerite riddled the strata below the lava fields. The lateral extent of the dolerites is indeed far wider than that of the basalt flows, many of which have been removed by denudation. Of these dolerite regions, that of the Karroo in the Union of South Africa is by far the most interesting. It is in the main a sparsely populated area owing to low rainfall, but these same conditions which hinder agriculture have resulted in magnificent natural exposures of unweathered rock. In

recent years, work on the railway and National Road systems of the country has added a number of excellent artificial sections. Only in the Transkei and Natal is the climate sufficiently wet to produce anything like a blanket of weathered material, and even there the more resistant sheets form bold escarpments, while excellent exposures may be seen along shore sections. The Karroo dolerites of the Union form, then, a hypabyssal province repaying detailed study, particularly as the thicker intrusions show interesting differentiation phenomena and also a well-marked and highly characteristic tendency to mobilize and metasomatize the associated sedimentary strata. Furthermore, the relationship of the dolerites to the overlying lava plateau is well displayed along the escarpments of Basutoland and the Stormberg region.

The present study is concerned principally with the dolerites injected into the strata of the Karroo System in the Union of South Africa. These intrusions cannot, of course, be considered apart from the accompanying volcanic episodes of the Stormberg and Lebombo, or from dikes and sills penetrating older strata, or Karroo beds from areas outside the Union. The main injection, however, is connected with the structural unit of the Karroo Basin, and the original work of the authors has been concentrated within this basin.

PREVIOUS WORK

The early petrological studies of Cohen (1887) on the Karroo dolerites were in advance of their time but have been overshadowed by the later investigations of Rogers and Du Toit.

Whatever progress may be made in the study of the Karroo dolerites, the pioneer work of Rogers and Du Toit will always be outstanding. It was carried out when transport was difficult, and in districts where accurate maps were only rarely available; yet they covered a vast area and assembled a great mass of trustworthy data covering both field relations and petrography. Much of this early work, as well as that of Schwarz, is embodied in the Annual Reports of the Cape Geological Commission from 1897 until 1911, and thereafter in the publications of the Union Geological Survey. It covers most of the Cape Province, including the Transkei Territories, and is summarized by Rogers (1905a, Chapter VII) and by Rogers and Du Toit (1909b, Chapter IX). A more recent summary of great importance is that of Du Toit (1920). It includes a detailed study of the mechanism of intrusion of the magma and an assessment of its place in Trias-Lias vulcanicity, as well as a description of certain highly differentiated intrusions in the Transkei.

A most interesting paper by Daly and Barth (1930) gives a detailed petrographical description with chemical analyses of four rather similar specimens, collected during the Shaler Exhibition, from widely separated localities in the Cape Province and the Orange Free State. Although the authors thought these specimens were representative, they give, in fact, a somewhat false impression of uniformity in the suite, as can be confirmed by reference to the earlier, though less accessible, literature. The four excellent analyses and the deductions drawn from them are, however, a valuable contribution to the subject.

Among the most notable recent additions to the literature are the work of Mountain (1935; 1936; 1943; 1944; 1945) on rheomorphic and syntectic effects produced by the

dolerites in sediments of T.M.S.[1] and Karroo origin; the detailed account of Scholtz (1936) of the great differentiated masses of East Griqualand and Pondoland, including the mineragraphy of the associated copper-nickel ores; and the studies by Frankel (1942; 1943) of multiple intrusions in the Aliwal North district. A detailed petrological account of a sill in Natal (De Swardt and Murray, 1944) contains an interesting discussion of the pyroxene crystallization in the dolerites, while Blignaut and Furter (1940), in their memoir on the Vryheid-Paulpietersburg coal field, have made important observations on the detection of dolerite types by marked differences in the field relations.

A number of papers by the present authors have been published since 1940. They include attempts to arrive at a working classification of Karroo dolerites (Walker and Poldervaart, 1941b; Walker, 1942), studies of sills in the Calvinia division (Walker and Poldervaart, 1940; 1941a), of metasomatism (Walker and Poldervaart, 1942a), and petrological investigations of two curious abnormal masses in the Transkei (Poldervaart, 1944; 1946).

COURSE OF PRESENT INVESTIGATION

Systematic field work was commenced in 1940 and took the following form:

1940. Intensive examination of the dolerites of the Calvinia district in the Western Karroo, and a reconnaissance across the Karroo to the Natal coast.

1941. Continuation eastward of systematic collecting (particularly along the Great Escarpment) as far as Northern Transkei, and intensive examination of intrusions in the Kokstad district.

1946. Collecting traverses across the Karroo from Beaufort West throughout the Orange Free State to the Southern Transvaal.

During the collecting work approximately 15,500 car miles were covered, but considerable areas remained untouched. These gaps were bridged by material kindly lent from the Geological Survey and Natal Museum collections, while further specimens were obtained through the help of senior students and research students in this department. In this connection, the collections made by Dr. G. T. Lamont were particularly useful. Much additional material was obtained through the road and railway authorities. The area covered by means of material from all sources is indicated in Figure 1.

ACKNOWLEDGMENTS

Though war commitments slowed down and even stopped the work for several years, its completion was hastened by the willing co-operation of a considerable number of petrologists and engineers. This included much helpful discussion, in the course of which valuable unpublished data and new specimens were made available to the authors. In this connection they are particularly grateful to Mr. P. B. Ackermann of Potchefstroom University College, Professor W. N. Benson of Otago University, Dr. C. F. Davidson, Dr. J. J. Frankel of the Natal University, Mr. J. J.

[1] Table Mountain Sandstone, an abbreviation frequently used in this work.

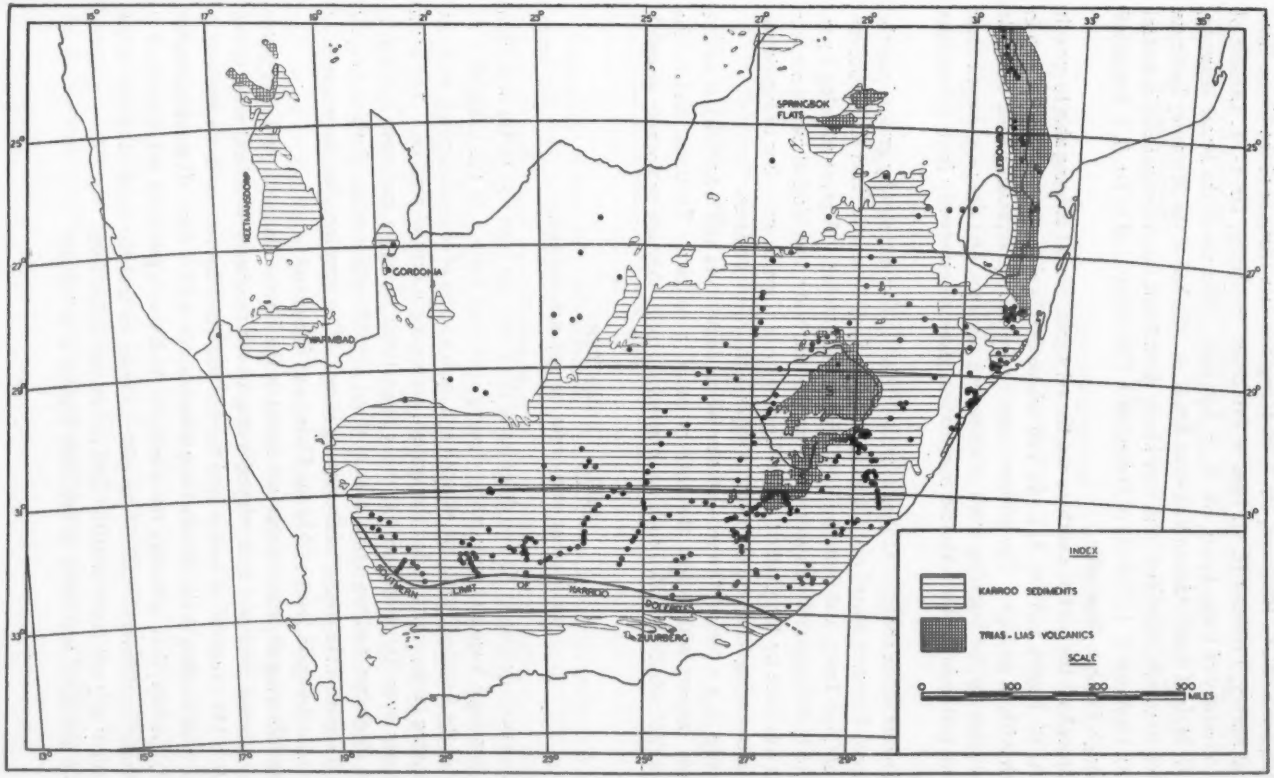


FIGURE 1.—*Distribution of Karroo System in the Union of South Africa*
Showing distribution of dolerites and lavas of Karroo age, examined by the authors.

Furter, Dr. G. N. G. Hamilton, Dr. S. H. Haughton, late Director of the Union Geological Survey, Professor H. H. Hess of Princeton University, Dr. G. T. Lamont of the University of Cape Town, Dr. R. F. Lawrence, Director of the Natal Museum, Mr. G. Mair of South African Railways, Dr. R. L. Mitchell of the Macaulay Institute for Soil Research, Aberdeen (who very kindly determined the trace elements in several rocks), Professor D. L. Scholtz of Stellenbosch University, and to Dr. S. I. Tomkeieff of King's College, Newcastle.

Throughout the work the authors had the very great benefit of being able to consult Dr. A. W. Rogers and Dr. A. L. du Toit whose deaths recently leave such a gap in South African geology. The authors cannot refrain from expressing their heartfelt admiration for the inspiring work of these two pioneers, and desire to record their sincere gratitude for the valuable information given so generously from this unique store.

The cost of the chemical analyses for this investigation was met by grants from the National Research Board and Council, from the Staff Research Fund of the University of Cape Town, and from the Royal Society of London. A grant covering a portion of travel expenses was received from the first body, and clerical assistance was made available by the Council of the University of Cape Town. For this generous and encouraging support, the authors express their indebtedness.

Lastly, it is a pleasure to recall the unfailing helpfulness and hospitality of all the Karroo farmers with whom the authors came in contact. To them, and to the "ysterklip" which provided such an interesting study, they are profoundly grateful.

FIELD CHARACTERS

GEOLOGICAL STRUCTURE OF THE KARROO BASIN

The main injection of Karroo dolerites in the Union has been into strata of Karroo age, forming a huge and very gently dipping structural basin. In this basin all divisions of the Karroo System are exposed and form a series of concentric rings, with the Stormberg lavas, which ended the system, occupying a central position. The basin measures over 800 miles and 400 miles respectively along its east-west and north-south diameters, and has a total area of about 220,000 square miles, of which roughly 10,000 square miles consist of Stormberg lavas. This area includes the volcanics of the Lebombo chain lying within the Union and Swaziland.

The following sequence of strata is found in the Karroo Basin of the Cape, where the maximum thickness is developed owing to the proximity of a land mass to the south. The existence of such a land mass is confirmed by the northerly inclination of the false bedding in the Beaufort sandstones south of 32°S. Lat. It must be emphasized, however, that, although the maximum thickness is generally in the south, it is developed in different localities for each division, so that the total thickness of the system at any one point probably does not exceed 20,000 feet.

The geological succession in Southern Karroo is as follows:

			Thickness in feet	European equivalent
	Unconformity			Rhaetic Lias
Karroo System	Stormberg Series	Drakensberg basalts	4,500	Triassic
		Cave Sandstone	800	
		Red Beds	1,600	
		Molteno Beds	2,000	
	Beaufort Series	Upper (Burghersdorp)	2,000	
		Middle	1,000	
		Lower	6,000	
	Ecca Series	Upper Middle Lower	6,000	Permian
	Dwyka Series	Upper shales	650	
		Tillite	2,500	
		Lower shales	750	
			27,800	Carboniferous
Cape System	Developed in Cape Folded Ranges	Witteberg Series	2,500	Devonian
		Bokkeveld Series	2,500	
		Table Mountain Series	5,000	
			10,000	
	Unconformity			

Throughout Karroo time deposition of fresh-water and terrestrial sediments continued quietly in this vast interior basin with only local interruptions. The glacial climate of Dwyka time was followed by conditions which varied between lacustrine, swampy, and desert. Toward the end of Karroo time a period of general aridity set in, and the Drakensberg lavas were poured out over the fine-grained aeolian deposits of the Cave Sandstone. The gentle synclinal structure of the Karroo Basin is due to warping, but to the south and west its edges have been crumpled by the great orogenic movements acting from these directions. The Cape Foldings, as these movements are called, were initiated in Witteberg time but reached their climax during the deposition of the Molteno Beds. Those to the west (the Cedarbergen) are the earlier and consist of open folds, but in the Southern Ranges (the Langebergen and Zwartbergen) there has been more severe crumpling, including overfolding and thrusting. Strata up to the Lower Beaufort division are affected. Small areas of Karroo strata have been preserved by mid-Cretaceous downfaulting in the heart of the Southern Ranges and bear witness to a former extension to the south of the basin of deposition.

To the east the Karroo Basin is bounded by the sea between 33° 20′ and 31° 30′ S. Lat. North of this latitude it is folded down by an easterly dipping monoclinal flexure of Liassic age, giving rise in the north to the Lebombo chain. In Natal and Zululand the hinge of the monocline has been worn away, exposing pre-Karroo strata.

North of the main basin numerous outliers of Karroo strata indicate a much greater original extent. Apart from their occurrence in the main Karroo Basin, strata of Karroo age, associated in many cases with basic lavas and intrusives, are found at numerous widely scattered localities in the Africa continent (Fig. 2).

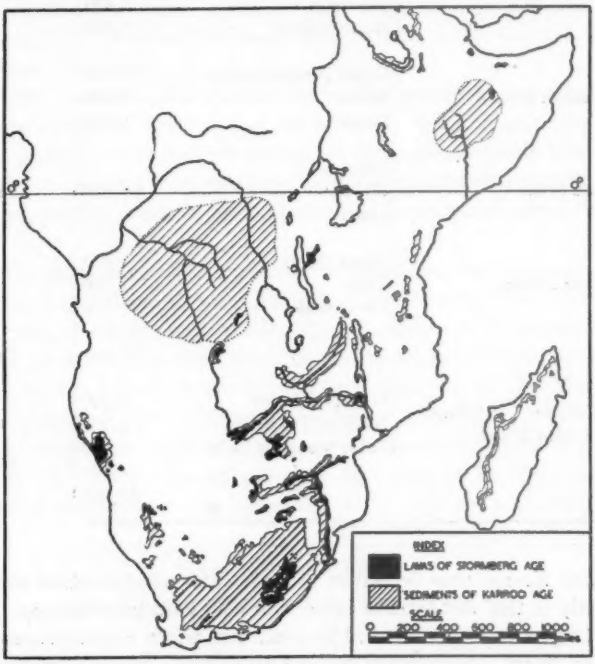


FIGURE 2.—*Distribution of Karroo System in Africa*
Showing distribution of sediments and lavas of Karroo age. (After Dixey, 1936).

DISTRIBUTION OF STORMBERG LAVAS AND SUMMARY OF THEIR FIELD CHARACTERS[2]

Drakensberg lavas.—The Drakensberg lavas have a restricted distribution in the center of the Karroo Basin, where they occupy roughly 10,000 square miles. They form the great block of Basutoland and attain there a maximum thickness of at least 4500 feet, reaching a height of 11,200 feet at Champagne Castle (Pl. 1, fig. 1). A subsidiary development is found immediately south of Basutoland in the direction of Barkly East in the Cape Province, while numerous small outliers occur west and north of the main plateau.

These lavas formerly occupied a far wider area, for fragments of identical volcanic rocks have been preserved in post-Karroo Kimberlite pipes as far west as Jagersfontein and even Prieska. Similar lavas of demonstrable pre-Cretaceous age cap

[2] The term "Stormberg", when applied to lavas and sediments, refers solely to their position in the geological time scale.

Stormberg sediments in the downfaulted Zuurberg region north of Port Elizabeth and within the Cape Folded Belt. In all the above cases basalt was the only type of lava erupted.

The Drakensberg lavas invariably form the highest portions of the landscapes where they are found, and their base never falls much below 5000 feet. The lower flows may show intercalations of Stormberg sediments, or of tuffs and agglomerates, but the bulk of the plateau is built up of an uninterrupted succession of basaltic lava flows ranging from 1 to 150 feet in thickness. These flows must have been highly mobile and poured out on a surface of low relief, for individual members may cover areas of over 200 square miles. The absence of boles or, indeed, of any signs of weathering of the upper surfaces of flows proves that the volcanic outpourings continued without prolonged pauses. A feature of the Drakensberg lavas is the occurrence of vertical pipe amygdules in the lower parts of the flows. These are filled with calcite, chalcedony, or zeolites and indicate a slow rise of gases, given out by the top of a cooling flow through liquid lava above. The tops of the flows are thoroughly vesicular, but in the thicker examples the central portions are dense and compact.

Although vents of central type are common in the Drakensberg lavas, they are mainly ash-filled and confined to the lower part of the succession. It is thought that the great plateau was built up as the result of fissure eruptions, for dikes of fine-grained dolerite may be traced upward through more than 3000 feet of basalts, and doubtless acted as feeders.

Zuurberg lavas.—The volcanic belt of the Zuurberg shows the same characters and is associated with tuffs and upper Stormberg sediments. It is of interest as showing that the Cape Foldings had been completed by Liassic times, for vulcanicity of this type is associated with tension rather than orogenic compression.

Lebombo lavas.—The lavas of the Lebombo belt, which outcrop along a north-south line between 23° and 29° S. Lat. show marked differences from those of Basutoland, although the two outpourings were practically contemporaneous. They are situated along a monoclinal flexure, which was actually initiated during the vulcanicity, and include highly differentiated types such as rhyolite and limburgite, in addition to basalts of normal Stormberg type. The eruption of the acid lavas was of an explosive nature with the production of layers of volcanic ash.

Activity commenced with the outpouring of basalts and limburgites in the region immediately west of the present range. According to Du Toit (1929b), this was accompanied by rifting and slight subsidence, lava welling out from the resulting fissures. Du Toit considers that a magmatic wedge was driven into the crust along a north-south line of weakness on the site of the main range, and that the rhyolites which now form its crest represent the accumulation of acid material toward the top of this wedge. Warping meanwhile maintained a general eastward slope of the whole belt, which now dips 5°–20° E., with still greater dips recorded near the axis of flexure. The rhyolitic eruptions were followed by a second series of basalts, which ended the sequence. The activity included the injection of large bodies of granophyre into the heart of the range.

The Lebombo lavas, by virtue of their mode of extrusion, accumulated to an immense thickness, estimated as at least 27,000 feet. Nevertheless, they form a rela-

tively insignificant feature in the landscape when compared with the majestic pile of Basutoland, where the thickness of lavas is only 4500 feet. This is partly due to their easterly dip, but also to climatic differences. The humid regions of the Kruger National Park and the adjacent Portuguese territory are covered with bush and weathered material, so that exposures are relatively poor, not only of the lower basic lavas, but of the underlying Stormberg sediments, in which the Forest Sandstone and Marl in the north take the place of the Cave Sandstone and Red Beds. The high ground of the range consists of the acid volcanics, which are characterized by flow banding and often seem to be much altered. The basic lavas, on the other hand, are often very fresh, in spite of a superficial appearance of weathering. Pipe amygdules, similar to those of the Basutoland block, are found in some of the flows. The limburgites may be distinguished from the normal basalts by their dense black appearance on fracture and by their high specific gravity.

The Liassic age of the sequence is proved by the presence of a sedimentary intercalation, containing the fossil Otozamites (Rennie, 1937).

The other outlying Karroo volcancis shown in Figure 2 have been less studied than the areas considered above and are not dealt with in this paper. The most important is that of the Springbok Flats.

GENERAL DISTRIBUTION OF KARROO DOLERITES

Outside the Cape Folded Belt the strata of the Karroo Basin are riddled with dolerite dikes and sheets, intruded contemporaneously with, or immediately following, the eruption of the Drakensberg lavas. The absence of Karroo dolerites from the Cape Folded Belt was commented upon by Rogers (1905a, p. 25), and their southern limit is marked on Figure 1. North of this line dolerites are found in profusion in all divisions of the system, though they show a tendency to extend laterally at certain definite horizons. Even in outliers north of the main basin dolerites abound, and the intrusive suite extends far beyond the Union into the Karroo areas of Warmbad, Keetmanshoop, and Gordonia (Fig. 1). To use the words of Du Toit (1920, p. 3),

> "Speaking roughly, the Karroo System between the twenty sixth and thirty-third parallels is riddled by dolerite, yet most curiously . . . the intervening areas of older strata from which the Karroo beds have been stripped (even when now forming small inliers), are rarely penetrated, and then only by minor intrusions".

Again (1920, p. 2),

> "The actual total area in South Africa within which the strata invaded are penetrated by these injections to such an extent that within the regions affected one can never get out of immediate sight of dolerite, is fully 220,000 square miles . . . while if allowances be made for the effect of erosion, and the outlying tracts are connected, the extent is increased by more than one half again".

Only those dolerites which are intrusive into the strata of the Karroo Basin, or the rocks immediately below it, are considered here.

Although the Karroo dolerites proper appear to be absent from the Cape Folded Belt, dolerite dikes of unknown age are abundant in the pre-Cape strata and granites of the Western Coastal region. A few of them penetrate the lowest beds of the T.M.S. They rarely exceed 100 yards in length and are of finer grain than the average Karroo

FIGURE 1. DRAKENSBERG LAVAS, MAIN GORGE, DRAKENSBERG NATIONAL PARK (NATAL)
Cave sandstone in foreground. (Photo South African Railways.)

FIGURE 2. EXECUTION ROCK, PORT ST. JOHNS (PONDOLAND)
(Photo South African Railways.)

LAVA AND SILL ESCARPMENTS

FIGURE 1. THE GREAT ESCARPMENT AT PERDEKLOOF (CALVINIA)
The upper part is formed by the Perdekloof dolerite sill, with Ecca shales and siltstones farther down.

FIGURE 2. DOLERITE DIKE IN THE MOLTENO PASS (BEAUFORT WEST)
The dike is inclined and has baked the adjacent Beaufort sediments.

FIGURE 3. THE THREE SISTERS (VICTORIA WEST)
The escarpments are formed by Beaufort siltstone, capped by dolerite. (Photo South African Railways.)

dolerite, from which they also differ in mineralogy and chemical composition. A full account of this dike suite has been given by Nell and Brink (1944).

GENERAL OBSERVATIONS ON FIELD RELATIONS OF KARROO DOLERITES

The Karroo strata into which the dolerites have been injected consist in the main of massive, feldspathic sandstones or siltstones, interstratified with thinly bedded or laminated mudstones and shales. As in other regions, where the intrusive magma and the sediments have the same characteristics—*e.g.*, in the Carboniferous of Central Scotland—sills are very prominent along the argillaceous bands, particularly in their lowest portions, while dikes invariably form relatively inconspicuous features in the landscape. Nevertheless, the Karroo intrusions show a greater variety of habit than is common. The transgressive form of the sheets, for instance, is often very marked, all gradations being seen between dikes and truly concordant sills.

The salient field characters of the Karroo dolerites may be summarised as follows: (1) absence of dolerites from strata of Folded Belt, (2) rise of sill horizon to the south-east in the Western Karroo, (3) abundance of dolerite sheets in Lower Beaufort strata and relative scarcity in the Dwyka tillite, (4) marked transgressive tendency of sheets, (5) abundance of ring-shaped intrusions in Central Karroo, due to basin-shaped transgressions of sills, (6) abundance of table mountains, capped by flat-lying dolerite sheets on Upper Karroo, (7) absence of tilting of strata by intrusions, (8) general inconspicuous appearance of dikes in the landscape, (9) increase eastward in size of intrusions (accompanied by tendency to develop differentiated types) and occurrence of a focus of intrusive activity in the Northern Transkei, (10) occurrence of bell-jar intrusions in, and west of, the focal region, (11) strong evidence of a sequence of injections from top to bottom in the focal region, and (12) extremely active behavior of magma toward the associated sediments at all horizons.

HABIT OF INTRUSIONS

Sills.—It has been remarked already that the Karroo dolerites show every gradation between concordant sheets and dikes (*e.g.*, Bulthouders Bank, Fig. 3). A great many intrusions are reasonably concordant and may thus be termed sills. Since the strata into which they have been injected are almost horizontal, the more nearly concordant sills have extensive outcrops and attain great prominence in the landscape.

Rogers (1905a, p. 252–255) has described several such sills from the Western Karroo. In every case, however, the intrusion cuts across the bedding to a certain extent as, for instance, the sill mentioned by Du Toit (1920, p. 6) from the extreme Western Karroo, which changes its horizon 600 feet in 54 miles. Where denudation has not dissected the sill to any great extent, an uplying boulder-strewn plateau, bounded by vertical and rudely columnar palisades, is found—*e.g.*, at Execution Rock in the Transkei (Pl. 1, fig. 2), and at Perdekloof, Calvinia district (Pl. 2, fig. 1). Prolonged erosion gives rise to the typical "table mountains" of the Upper Karroo, and in later stages to "koppies" of the "theebus" or "koffiebus" type (Pl. 2, fig. 3). The original extent of some of these sills has been very great. Du Toit (1939, p. 327) has estimated the area of one sill in the Hopetown division as at least 5000 square miles.

Little systematic work has been carried out on the sills of the Orange Free State, but traverses of this province indicate that concordant sills of extensive outcrop are abundant.

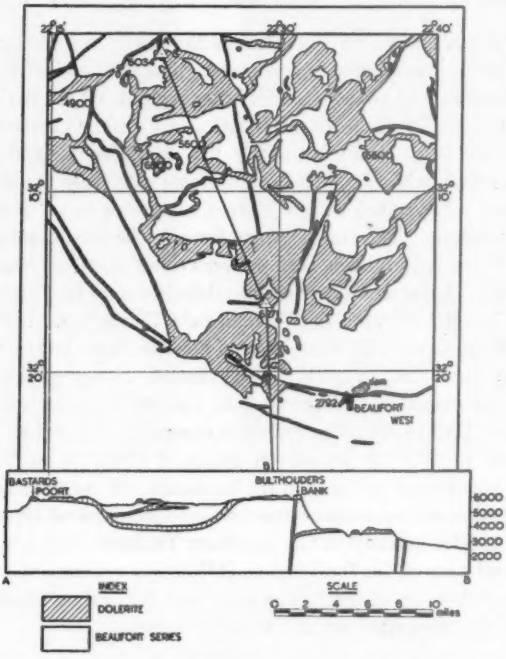


FIGURE 3.—*Map and section of Bulthouders Bank sheet, Beaufort West*
(After Rogers and Schwarz, 1911).

Where the strata are thinly bedded, as in the southern Ecca, the transgressions may be inconspicuous, but in the case of massive, arenaceous beds alternating with argillaceous beds, as in the Beaufort Series, the transgressions are often "stepped".

There is the usual tendency for the ascending magma to spread out laterally along argillaceous horizons of easy parting, more especially in the lower portions immediately above massive arenaceous strata with prominent vertical joints. Du Toit (1920, p. 5) has suggested that the lateral spread may be accentuated by the distillation of water vapor and other volatiles from the heated sediments. These gases would have a splitting effect on the argillaceous strata, which would facilitate their penetration laterally by the ascending magma.

Some of the concordant sills develop conspicuous undulations or pass into markedly transgressive sheets.

The sills range in thickness between 1 foot to over 1000 feet. Those in the Western Karroo do not seem to exceed 500 feet, a maximum reached at Hangnest (Walker and Poldervaart, 1941a) and Bulthouders Bank at Beaufort West (Rogers and Du Toit,

1909b, p. 264). In the Eastern Cape Province and Transkei much greater dimensions are obtained, several sills having a thickness of over 1000 feet—*e.g.*, at Mount Prospect near Libode.

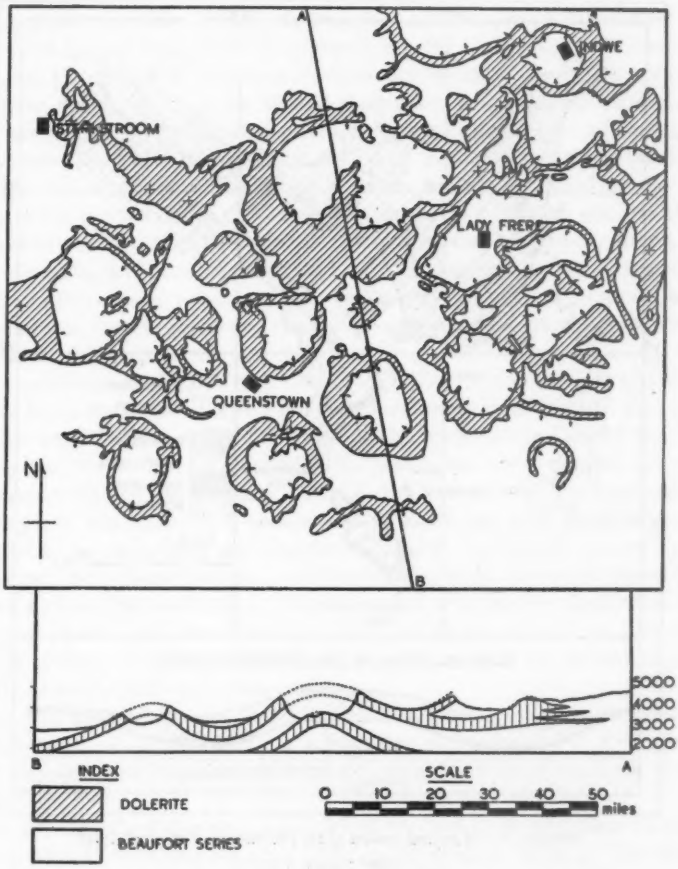


FIGURE 4.—*Map and section of the Glen Grey sheet, Queenstown*
(After Rogers and du Toit, 1909b).

Curved sheets.—The undulating Karroo sheets are often basin-shaped with quaquaversal dip, which may be gentle or comparatively steep. If sediments occur above the sheet in the center of the basin, the result is a ring-shaped and sometimes almost circular outcrop of dolerite (Fig. 4). At Bongola near Queenstown, a dam has been constructed across a break in the ring to form an excellent storage reservoir.

Such undulating intrusions attain their greatest development in the Queenstown-Glen Grey area but have been recorded from the divisions of Fraserburg, Victoria

West, Britstown, de Aar, and Umtata, as well as from East Griqualand. Their flatter portions are up to 1500–2000 feet thick in the Andriesbergen and Wildschutsberg in the Queenstown area.

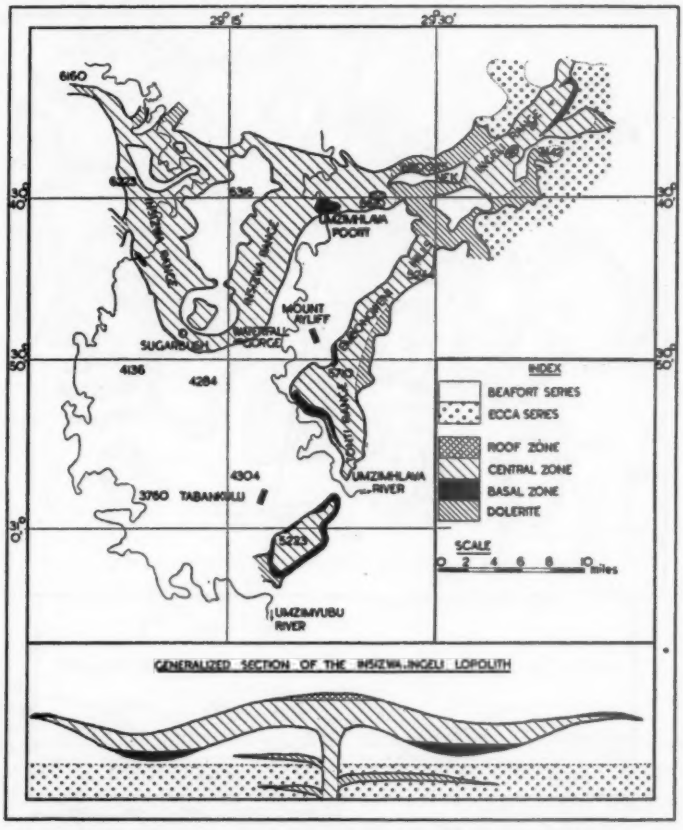


FIGURE 5.—*Map and section of the Insizwa sheet, Mount Ayliff*
(After Scholtz, 1936).

Still thicker are the huge curved sheets of Insizwa, Ingeli, and Mount Ayliff in East Griqualand and Pondoland, which exceed 3000 feet. These masses thin in their upward undulations but thicken toward the intervening troughs (Fig. 5) in which crystallization differentiation has taken place with settling of olivine crystals and production of picrites and peridotites.

Dikes.—Apart from undulatory sheets and from occasional transgressions of sills which are elsewhere concordant, numerous sheets remain transgressive for long distances. The steeper of these grade into dikes (Pl. 3, fig. 3), and in some cases have

been referred to as such—*e.g.*, the intrusion at Beaufort West dam (Rogers, 1905a, p. 256). The inclined sheets are 600 feet thick at the Glen, 25 miles north of Graaff Reinet (Walker and Poldervaart, 1942b, p. 58), but this is unusual, and the majority are much thinner.

Most of the true dikes—that is, the intrusions which approach the vertical in habit—are much thinner than the inclined sheets, ranging from 8 feet to 28 feet in width (Pl. 3, fig. 1). However, they may be more than 40 miles long. These dikes seem, in general, to be later than the sill and sheet phase and often cut these intrusions. Dikes are abundant throughout the Central Karroo but relatively thin and sparse in the Western Karroo and in the coastal parts of the Transkei and Natal (Du Toit, 1920, p. 13). Their abundance is, to a certain extent, controlled by the division planes of the injected strata. Dikes, for instance, may be sparse, and sills abundant in the thinly bedded Ecca, while in the adjacent Lower Beaufort sandstone, with its massive bedding and prominent vertical joints, the reverse is the case. In pre-Karroo formations there seems to be very little tendency for the magma to spread laterally, and nearly all the intrusions are thin dikes. The only exceptions are a few sills in the T.M.S. of Natal and the Transkei.

The main trend of the true dike phase is WNW. in the focal region of Basutoland and the Northern Transkei, with a less prominent set at right angles. In the Stormberg the trend is NW., in the Lebombo due N., while in the Central and Western Karroo it is NNW.—that is, parallel to the post-Nama and Cedarberg foldings.

Abnormal dikes.—The two gap dikes of the Transkei reach a maximum width of 320 yards and maintain a parallel course E.–W. for over 90 miles from near Cathcart to the sea. They are exceptional, not only in their dimensions but also in their aspect as landscape features. Furthermore, the dominant rock type is a coarse-grained and relatively acid type. They were first described by Rogers and Schwarz (1901b, p. 64–65; 1904), but Mountain (1943) has given a recent and comprehensive account of them, including micrometric and chemical data, while the senior author (1943) has dealt briefly with their origin.

The chief point about the field relations of these intrusions is their liability to weather out more rapidly than the surrounding sediments and so form prominent east-west gaps across the landscape. Their outcrop is marked by scattered exfoliating boulders.

The dolerite sills of this region are pierced by even more acid dikes as well as by the gap dikes (Rogers and Schwarz, 1901b, p. 65; Rogers, 1905a, p. 261; Mountain, 1944, p. 116). At Kabonqaba mouth a narrow dike is figured by Rogers, while at Gqunque a much larger mass is considered by Rogers and Schwarz to be a short dike, though Mountain believes it to be a capping of the adjacent Manubi dolerite sheet.

Elephant's Head dike is a large multiple dikelike intrusion (at least 25 miles long and 50–100 yards wide) running E.–W. across the Matatiele division into Basutoland. It has been described by Du Toit (1929a) and, more recently, in greater detail by the junior author (1944). This dike has an intermittent outcrop and is definitely headed. There is also a strong suggestion that it may be floored as well, for picrites occur along the lowest exposures. From their relations to the normal dolerite, the

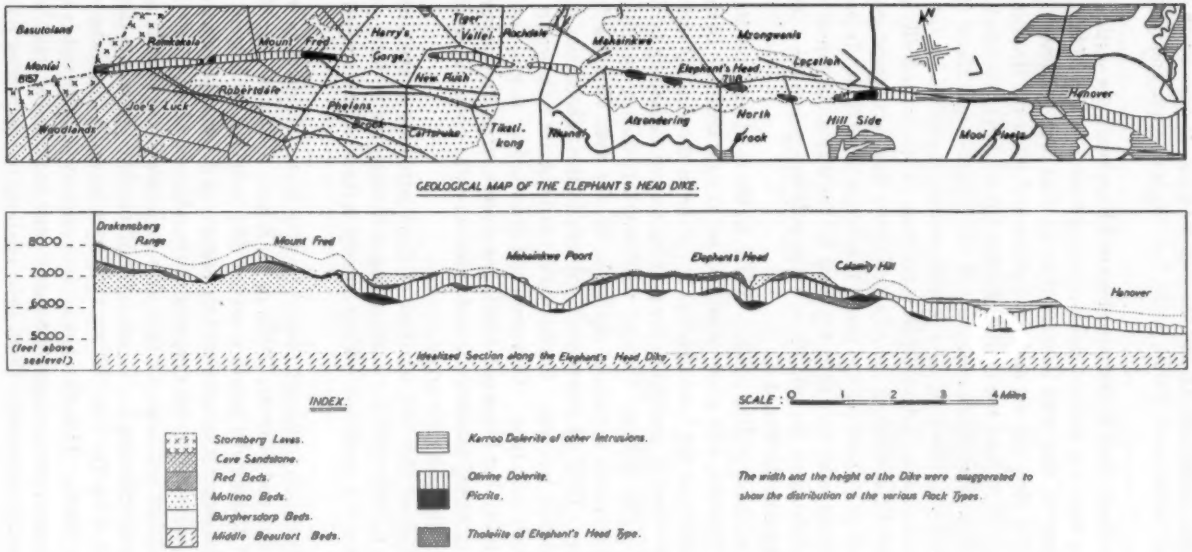
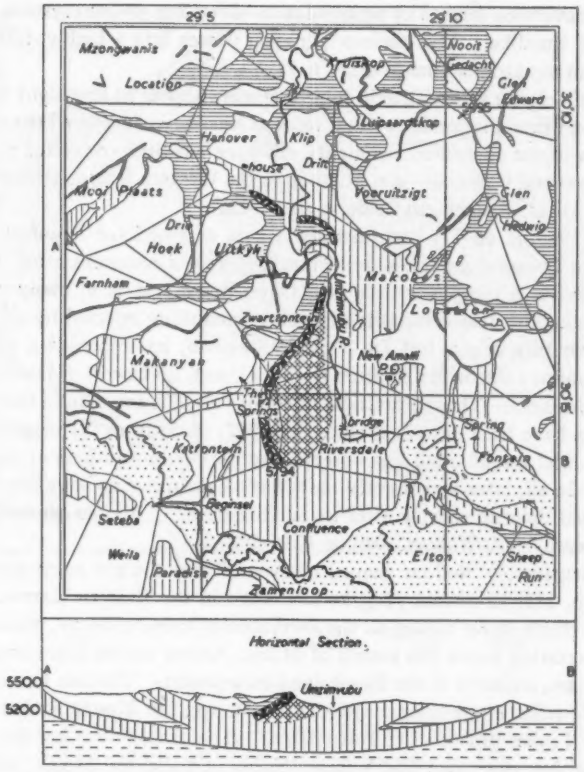


FIGURE 6.—*Map and section of the Elephant's Head dike, Matatiele*
(After Poldervaart, 1944).



MAP OF THE NEW AMALFI SHEET.

LEGEND.

SCALE: 0 1 2 3 Miles

Soil of the Cedarville Flats.

Upper Beaufort or Burgersdorp Beds.

Middle Beaufort Beds.

Dolerites with small olivines.

Metasomatic Granophyres.

Iron-rich Dolerites.

Olivine-Dolerites.

FIGURE 7.—*Map and section of the New Amalfi sheet, Matatiele*
(After Poldervaart, 1944).

picrites must have been formed by accumulation of settling olivine crystals, probably on the floor of the dike. This curious intrusion passes into a highly differentiated sill with an acid top at New Amalfi (Figs. 6, 7).

A number of dolerite sills with acid tops or centers appear at first sight to be composite, but closer examination shows that some at least have developed the acid phase by transfusion of the associated sediments; siltstone has been converted to a granophyre rock in several instances—*e.g.*, Rietkop in the Western Karroo (Rogers and Du Toit, 1903, p. 41–42; Walker and Poldervaart, 1942a).

Mountain (1943, p. 66–71) has described a core of mobilized Beaufort sediment from one of the Transkei gap dikes, and Frankel (personal communication) has found similar relations in a composite dike near Queenstown. It is in many cases very difficult to distinguish true composite intrusions from those injected by rheomorphic veins of sedimentary origin, but Du Toit (1920, p. 16) has recorded a narrow and apparently genuine composite dike from Oorlog's Poort, northwest of Dordrecht.

Multiple intrusions.—Though multiple injections are not common in the Karroo, a few such dikes have been described (Frankel, 1942; 1943) from the neighborhood of Aliwal North, and a similar sill was observed by the present authors at Lovani near Steynsburg. In all, later and slightly more basic basaltic veins have been intruded into the normal dolerite shortly after its solidification. This phenomenon appears to be widespread. (*Cf.* Walker, 1940, p. 1086–1087.)

Bell-Jar intrusions.—"Bell-jar shaped intrusions" was the apt name given by Du Toit (1905a, p. 131) to certain pluglike intrusions in the Eastern Karroo. Two of them lie near Bird's River Siding on the Sterkstroom-Indwe railway, while the third and most interesting forms the massif of Mount Arthur at the boundary of Natal and the Transkei, adjacent to the Basutoland escarpment. The two former measure 6 miles and 1½ miles along their major diameters, and the Mount Arthur mass is 7 miles long by 4 miles broad. Du Toit (1905a; 1911b, p. 132–134) has described the Stafelberg Vlei intrusion near Bird's River Siding and also the Mount Arthur mass (1913, p. 96) which has recently been dealt with in greater detail by the junior author (1946). The essential features of all three are the isolation and foundering of a cylindrical block of sediments through the ascent of dolerite magma along a ring fracture (Fig. 8).

These masses show many of the characteristics of the well-known "cauldron subsidences", which are so striking in the Western Highlands of Scotland (Clough, Maufe, and Bailey, 1909, p. 61), but differ from them by causing arching and fracturing of the roof. The foundered plug of sediments is nowhere exposed, but the actual roof is seen at Mount Arthur, where detached blocks of sediments, belonging to the Cave Sandstone, Red Beds, and Molteno Beds, and dipping at random, have sunk over 1000 feet in the magma. Possibly the ascending magma may have actually broken through to the surface, forming a caldera. Lateral offshoots of dolerite transgress at varying inclinations from the main plug, like cone sheets.

Laccolith.—The only laccolithic mass in the Karroo is the small knob of Jackal's Kop, 5 miles WNW. of Dordrecht (Du Toit, 1920, p. 13). It has arched up the overlying strata but is perhaps more correctly termed a phacolith, since it occurs on a sharp monoclinal flexure.

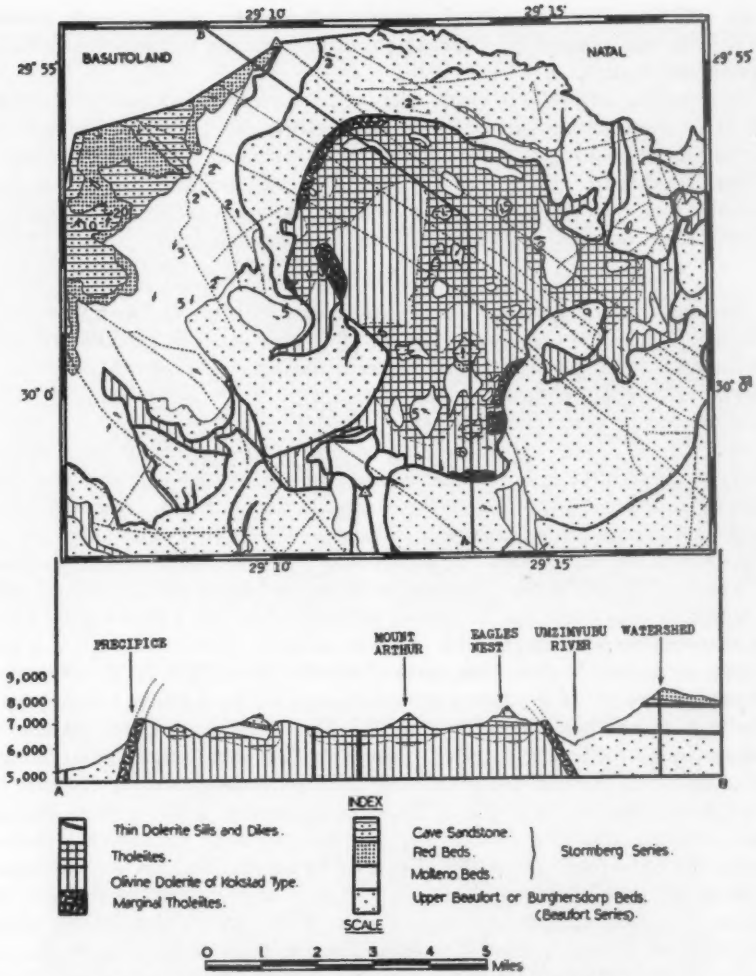


FIGURE 8.—*Map and section of the Mount Arthur complex, Kokstad*
(After Poldervaart, 1946).

Pegmatitic schlieren and residual injections.—A few of the thicker intrusions show schlieren of coarse dolerite pegmatite which do not exceed 2 or 3 feet across and are usually found toward the top of sills, and roughly parallel to the contact (Pl. 4, fig. 1). They tend to develop in the neighborhood of xenolithic blocks (*e.g.*, at Alewyn's Gat, Fig. 37).

The best exposures are in artificial rock cuttings, and schlieren may be more common than is indicated by natural outcrops, where they may be masked by the weathered

crust. All the main types of dolerite contain such schlieren, which strongly resemble the dolerite pegmatites of other tholeiitic provinces and are considered the result of crystal fractionation.

Thin irregular acid veins in the dolerite (*e.g.*, at Downes Mountain and Hangnest, Calvinia district) are thought by the present authors (1940, p. 165–166) to represent late injections of a highly differentiated partial magma. They are considerably fresher than the acid types produced by the metasomatism of Karroo sediments, have no visible connection with sedimentary material, and again find numerous analogues in other tholeiitic provinces, where the magma is relatively inert.

JOINTING AND MODE OF WEATHERING

The general aspect of a dolerite outcrop seems to depend on (1) the arrangement of the division planes in the dolerite, (2) the inclination of the intrusion, and (3) the liability of the rock to alteration by subaerial weathering. Factors (1) and (2) are related, for the dolerite types which are particularly prone to alteration differ in their jointing from the more resistant types.

On the whole olivine dolerites seem to be more resistant than the relatively acid types, which carry abundant columnar bronzite or magnesian pigeonite at the expense of the orthosilicate. The jointing of olivine dolerites is generally rudely columnar, so that many sills form prominent palisades, along the crest of the Great Escarpment or crowning table mountains and conical hills of the "spitskop", "theebus", or "koffiebus" type (Fig. 9). The rude polygonal columns of such sills are traversed by numerous cross joints, and the thinner sills seem to develop a more regular columnar structure than the larger ones. Curved columnar joints are rare, but a good example occurs near Molteno Pass, north of Beaufort West (Pl. 4, fig. 2). Sometimes a columnar "krans" of Beaufort sandstone is crowned by a gently rounded cap of dolerite, as at the Three Sisters (Pl. 2, fig. 3). There is indeed a marked tendency for Beaufort sandstone to form steep escarpments showing prominent columnar jointing.

Though gently inclined sheets show columnar escarpments, steeper dips give rise to a different type of topography. The flat-lying parts of olivine dolerite sills may show columnar palisades, whereas the dip slopes of more transgressive portions develop exfoliating slabs (Pl. 4, fig. 3), relieved by isolated boulders in small koppies of disordered joint blocks (Pl. 4, fig. 4). Lines of such koppies connected to one another frequently mark the outcrop of dikes or of the curved and undulating sheets. The joint blocks may be rounded by exfoliation, or unweathered and angular, and some appear to have been split by insolation. More rarely the piles of loose blocks give place to solid walls of masonry, known as "mural weathering". This rectangular jointing is developed to perfection in the Valley of Desolation near Graaff Reinet (Pl. 5, fig. 1). The dolerite is dominantly reddish brown on outcrop, though many boulders in the more arid areas are coated with a black desert varnish. Reddish brown is also the usual color of dolerite soil, except in the most humid regions, where brick-red laterite or black clay soils are developed.

Over the whole of the Karroo coarse bronzite dolerites of the Hangnest type characteristically weather rapidly. This applies also to the associated dolerites of the Downes Mountain type, which form the upper part of many Hangnest type sills.

FIGURE 1. NARROW DIKE IN BEAUFORT, FRERE (NATAL)
(Photo E. S. W. Simpson).

FIGURE 2. JOINTING OF DOLERITE IN MOLTENO PASS (BEAUFORT WEST)

FIGURE 3. HIGHLY TRANSGRESSIVE SHEET IN THEEKLOOF (FRASERBURG)
Baking of adjacent Beaufort sediments has rendered them more resistant to weathering.
(Photo A. W. Rogers).

JOINTING AND WEATHERING OF KARROO INTRUSIONS

FIGURE 1. PEGMATITIC SCHLIEREN IN UPPER PART OF BULTHOUDERS BANK SHEET
Alewyn's Gat (Beaufort West).

FIGURE 2. JOINTING IN THIN DOLERITE SILL
Molteno Pass (Beaufort West).

FIGURE 3. EXFOLIATION IN DOLERITE
Molteno Pass (Beaufort West).

FIGURE 4. ISOLATED "KOPPIE" OF DOLERITE JOINT BLOCKS,
BEAUFORT WEST.

PEGMATITE SCHLIEREN AND WEATHERING
PHENOMENA IN SILLS

So readily do these types weather that they seldom form escarpments on the hillside, and their outcrop is marked only by exfoliating slabs with perhaps a few koppies of rounded joint blocks, representing the more resistant portions (Pl. 6, fig. 1). The contrast in the field between these dolerites (Matshongololo) and the columnar Zuinguin type of olivine dolerites (Perdekloof of the present authors) has been emphasized by Blignaut and Furter (1940, p. 50).

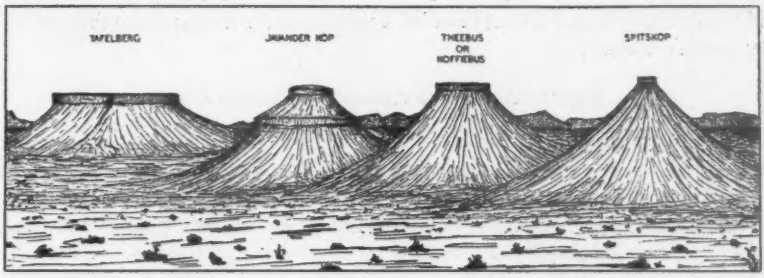


FIGURE 9.—*Sketch of the type of "koppies" produced in the Karroo by dolerite sills*

Where blocks of sediments have been stoped off by the dolerite magma, or where sedimentary partings occur in a sill, the resulting metamorphosed sediments form much more prominent features than the enveloping dolerite. At Rietkop, for instance (Pl. 5, fig. 2), a sedimentary parting of Lower Beaufort siltstone has been metasomatized to a hard granophyre which forms a prominent escarpment on the hillside, but the surrounding dolerite can be detected only at close quarters.

BEHAVIOR OF THE DOLERITE MAGMA TOWARD ASSOCIATED SEDIMENT

In nearly every case the dolerite magma shows marked chilling against the sediments into which it is injected, and in many places a tachylite selvage is developed. Nevertheless, the contact between sedimentary and igneous rock is sometimes gradational, and the Karroo province as a whole is characterized by the extreme activity of the magma toward the country rock. This is in marked contrast to other dolerite provinces of similar magma type, such as those of North England and Central Scotland, and of the Acadian region where the magma is virtually inert.

The marginal sediments generally show the effect of normal optalic metamorphism. Argillites, the most common contact rocks, are baked to dense hornstones with conchoidal fracture, while arenaceous beds are toughened and recrystallized. In sills metamorphism seldom extends far from the contact, but in dikes it may be considerable. In Theekloof (Du Toit, 1939, p. 260) the course of a dike is marked by two exceedingly prominent walls of hornfelsed sediment, each of which is thicker than the dolerite. Columnar jointing of a dike or sheet may extend for a considerable distance into the adjacent sediments (Pl. 2, fig. 2; Rogers, 1905a, Pls. XIII, XVI). Of some commercial importance is the anthracitization by Karroo dolerite sills of the Middle Ecca coals of Natal and the Transvaal. Blignaut and Furter (1940, p. 67)

have shown that olivine dolerites of the Zuinguin type bring about such changes for only one sixth of the distance altered by the later Matshongololo bronzite dolerite.

In rare instances the magma has actually fused the adjacent country rock. Du Toit (1904, p. 176–177) has described a case at Coldbrook, Barkly East, where a dolerite sill has not only fused the adjacent Cave Sandstone to a green pitchstone glass but has been injected by rheomorphic veins of the vitreous material. Similar instances from the neighborhood of Heilbron have been described by Young (1918), and Du Toit (1905a, p. 131) has recorded a strip of cordierite glass from the Stafelberg Vlei bell-jar intrusion.

MOBILIZATION AND METASOMATISM OF SEDIMENTS

The most striking field feature of the Karroo dolerite magma is its power of mobilizing and metasomatizing associated sediments. Rheomorphic veins of sedimentary material are everywhere common in the dolerites, and many can be traced either to the contact or to some parting or xenolithic inclusion. Karroo sediments of all ages, and even the underlying T.M.S., seem to be affected in this manner. The phenomenon is readily observable in the field, and it is rare to encounter an intrusion of any size which does not show some sign of it.

At Gugweni Location in Umzimkulu Karroo dolerite has reacted with the matrix of xenoliths of Dwyka tillite, leaving the pebbles unchanged, and at Doornkloof, southeast of Ladysmith, arkoses of Ecca age have been converted into rocks resembling microgranite, though they retain traces of false bedding (Du Toit, 1913, p. 91; 1920, p. 15).

Mountain (1935, p. 93–112) has described rheomorphic and syntectic phenomena at Coedmore quarries near Durban, where a Karroo dolerite sill has mobilized the adjacent T.M.S. and has been injected by the resulting rheomorphic veins. Reaction between dolerite and mobilized sediment has produced decomposed contaminated types. Mountain (1936, p. 248–253) has also described rheomorphic veins of argillaceous material in a dolerite from the East London neighborhood and has recorded numerous other cases of mobilization of Beaufort sediments (1944, p. 107–121). He (1943, p. 55–74) regards the greater part of the gap dikes of the Transkei, and practically all the acid igneous-looking material associated with the dolerites, as the product of syntexis.

The present authors (1942a) found, during a study of selected examples, that in two cases (Rietkop, Calvinia, and Alewyn's Gat, Beaufort West) relatively unaltered cores of siltstone remained embedded in the granophyric products, which are the result of its transfusion by dolerite. In both cases the junction between metasomatic granophyre and dolerite was sharp, but in the case of other, argillaceous sediments, gradational contacts were sometimes observed—*e.g.*, at the Glen, north of Graaff Reinet (Walker and Poldervaart, 1942b, p. 56–57). At Hangnest, Calvinia (Walker and Poldervaart, 1941a), proof of transfusion is not so complete. A large block of Beaufort siltstone has been partially stoped from the upper contact of a 500-foot sill and has a shell of chilled basalt, locally brecciated by rheomorphic material (Fig. 10). A detached fragment of this block deeper in the sill has been converted to a metasomatic granophyre, similar to those of Rietkop and Alewyn's Gat (Figs. 37, 38).

FIGURE 1. MURAL WEATHERING IN DOLERITE
Valley of Desolation (Graaff Reinet). (Photo South African Railways).

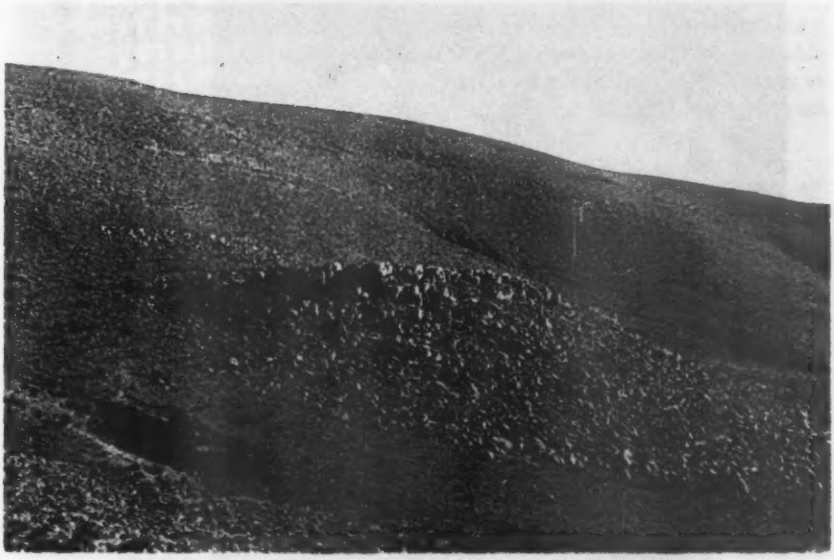


FIGURE 2. OUTCROP OF METASOMATIC GRANOPHYRE AT RIETKOP (CALVINIA)
Upper, coarse-grained granophyre forms an isolated band above the larger exposure of fine-grained granophyre. Above and below the granophyre are exposures of altered, crumbly dolerite.

CONTRASTED WEATHERING OF DOLERITE TYPES

FIGURE 1. DOLERITE SILL AT HANGNEST (CALVINIA)
Escarpment consists of Beaufort siltstone, with dolerite forming smoother slope. Metasomatic granophyre outcrops in immediate foreground, while stoped block of siltstone dips toward observer.

FIGURE 2. RHEOMORPHIC SILTSTONE VEIN IN DOLERITE, BEAUFORT WEST RESERVOIR
Vein has been pinched off and dislocated, due to jointing in dolerite while rheomorphic material was still plastic.

FIGURE 3. SILTSTONE XENOLITH WITH SEVERAL RHEOMORPHIC VEINS IN DOLERITE
Beaufort West reservoir.

XENOLITHS AND RHEOMORPHIC VEINS

Where siltstones and coarser arenaceous sediment are involved, the mineral changes brought about by metasomatism and mobilization are clearly visible in the field. The recrystallized rock generally has an igneous appearance, but the rheomorphic veins of argillaceous composition show no megascopic difference from the parent hornstones.

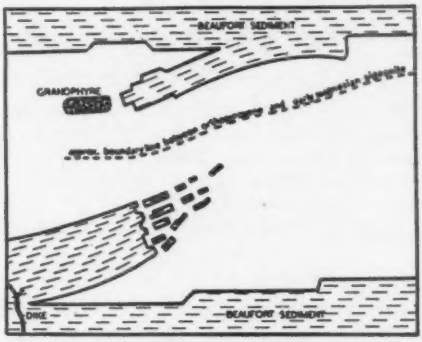


FIGURE 10.—*Section of the Hangnest sill, Calvinia*
(After Walker and Poldervaart, 1941a).

Some unusual syntectic granophyres have been reported (Du Toit, 1929a, p. 24–25) from the dolerite sill of New Amalfi (Matatiele) and have been described in detail by the junior author (1944, p. 107–112).

Though rheomorphism and transfusion at many of the above localities are striking in the field, many of the outcrops are inaccessible without motor transport. Notable exceptions are Coedmore quarries at Bellair, near Durban, and the inclined sheet at Beaufort West dam, which show excellent examples of these phenomena on a small scale.

PETROGRAPHY

CLASSIFICATION

In classifying the Karroo dolerites the authors have reduced the number of common types to four. Most of the Karroo igneous rocks can conveniently be classified according to this scheme, but some modifications and rarer types of local occurrence are dealt with separately.

The following are the main types throughout the Karroo:

(1) *Perdekloof Type*—type locality: Perdekloof (31° 47′ S. Lat., 19° 57′ E. Long.). Medium- to fine-grained ophitic olivine dolerites (Pl. 7, fig. 1).

Pyroxene is generally untwinned, strongly zonal augite, fringed by ferriferous pigeonite in the coarser examples. Olivine varies in amounts. Strongly zonal plagioclase occurs as slender laths 0.2–0.7 mm. long.

Chief characteristics are the ophitic texture, shown by a spotted appearance in the hand specimens, and the absence of orthopyroxene and columnar, magnesian pigeonite.

(2) *Blaauwkrans Type*—type locality: Blaauwkrans (31° 39′ S. Lat., 19° 43′ E. Long.).

Medium-grained, subophitic olivine dolerites and tholeiites (Pl. 7, fig. 2).

Augite frequently envelopes columnar cores of magnesian pigeonite. Twinning on (110) or (100) common. Olivine occurs in variable amounts or may be absent. Plagioclase builds zonal laths 0.6–0.8 mm. long; and rarely exceeds 50 per cent by volume.

Chief characteristics are the subophitic texture, the occurrence of columnar magnesian pigeonite, the absence of orthopyroxene, and the high color ratio.

(3) *Kokstad Type*—type locality: near Kokstad (30° 36′ S. Lat., 29° 29′ E. Long.).

Coarse-grained, subophitic olivine dolerites (Pl. 7, fig. 3).

Augite occurs with orthopyroxene or pigeonite. Olivine varies greatly in amounts. Plagioclase forms zonal laths 0.8–1.4 mm. long. Except in the olivine-rich modifications, plagioclase forms more than 50 per cent by volume. Micropegmatite may occur interstitially, but no mesostasis has been found in rocks of this type.

Chief characteristics are the coarse grain, the subophitic texture, and the low color index.

(4) *Hangnest Type*—type locality: Hangnest (31° 43′ S. Lat., 20° 2′ E. Long.).

Medium- to coarse-grained, subophitic bronzite dolerites (Pl. 7, fig. 4).

Augite occurs in addition to orthopyroxene, with or without pigeonite. Olivine is absent. Plagioclase builds zonal laths 0.6–1.2 mm. long.

Chief characteristics are the subophitic texture, absence of olivine, occurrence of magnesian orthopyroxene, and abundance of micropegmatite.

Table 1 shows the frequency, expressed as percentages, of the various types in the different parts of the Karroo Basin.

TABLE 1.—*Relative frequency of main dolerite types in Karroo*

Type	20–24°E. Long	24–28°E. Long.	Transkei	Natal	Whole Karroo
Perdekloof	23.5	46.9	5.5	26.7	28.4
Blaauwkrans	39.2	29.7	14.4	12.0	23.6
Kokstad	2.0	6.2	54.3	10.7	14.7
Hangnest	9.8	1.6	2.8	20.0	9.3
Other types	15.5	15.6	23.0	30.6	24.0

The following types are rarer:

(5) *Downes Mountain Type*—type locality: Downes Mountain (31° 32′ S. Lat., 19° 58′ E. Long.).

Medium- to coarse-grained, subophitic pigeonite dolerites (Pl. 7, fig. 5).

Columnar, magnesian pigeonite occurs in addition to augite. The two pyroxenes often form ragged, subophitic prisms with marked elongation and curvature. Both olivine and orthopyroxene are absent. Plagioclase forms zonal laths 0.6–1.2 mm. long. Characteristically the type occurs in the upper third of sills of Hangnest type dolerite.

Chief characteristics are the subophitic texture, absence of both olivine and orthopyroxene, abundance of micropegmatite, and its association with the Hangnest type.

(6) *Hanover Type*—type locality: Hanover (31° 5′ S. Lat., 24° 26′ E. Long.). Medium-grained, ophitic olivine dolerites, with large crystals of orthopyroxene 1.5–2.5 mm. long (Pl. 7, fig. 6).

Augite, with or without ferriferous pigeonite, occurs in addition to orthopyroxene. Iron-rich olivine is found in small quantities. Plagioclase builds slender laths 0.4–0.6 mm. long.

Chief characteristic is the presence of large orthopyroxene crystals, plainly visible in the hand specimens.

(7) *Marginal tholeiite*—type locality: Linton (29° 58′ S. Lat., 29° 12′ E. Long.). Medium-grained, iron-rich tholeiites (Pl. 8, fig. 1).

Slightly zoned, euhedral subcalcic ferroaugite occurs with or without pigeonite. Olivine may occur in small amounts. Moderately zonal plagioclase laths are 0.6–1.0 mm. long. Chlorophaeite often occurs as an accessory mineral.

Chief characteristics are the euhedral texture and the enrichment in iron and titania of these rocks.

(8) *Kranskop Type*—type locality: Kranskop (30° 36′ S. Lat., 26° 32′ E. Long.). Fine-grained, subophitic olivine dolerites (Frankel, 1942, Pl. II, fig. 1).

Zonal augite occurs in elongated crystals. Olivine is a prominent constituent (> 10 per cent by weight). Plagioclase builds laths of average length 0.2–0.4 mm. Picotite is a typical accessory.

Chief characteristics are the prominence of olivine and the occurrence of accessory picotite.

(9) *Picrite and Picrite-dolerite*—type localities: Naauwpoort (31° 24′ S. Lat., 25° 2′ E. Long.), and Calamity Hill (30° 7′ S. Lat., 29° 1′ E. Long.).

Coarse-grained, ophitic or poikilitic rocks, with more than 25 per cent by weight of olivine (Pl. 8, fig. 2).

Augite and orthopyroxene build ophitic plates. Plagioclase occurs either as small zonal laths 0.4–0.6 mm. long (*e.g.*, Naauwpoort), or builds large crystals 1.0–1.8 mm. long, poikilitically enclosing olivine (*e.g.*, Calamity Hill).

Chief characteristic is the high percentage of olivine.

(10) *Dolerite-pegmatite*—type locality: Alewyn's Gat (32° 14′ S. Lat., 22° 34′ E Long.).

Coarse-grained segregations occurring as veins or schlieren in dolerite intrusions (Pl. 8, fig. 3).

Augite, ferroaugite, and pigeonite may occur, but orthopyroxene is absent. Olivine may be found in small amounts. Tabular crystals of plagioclase are 2.0–4.0 mm. long. Micropegmatite occurs interstitially. With an increasing amount of micropegmatite the type becomes granophyric.

Chief characteristics are the extremely coarse grain, occurrence as veins or schlieren, absence of orthopyroxene, and abundance of micropegmatite.

(11) *Kentani Type*—type locality. Southern Kentani dike, Butterworth, Kentani Road (32° 26′ S. Lat., 28° 18′ E. Long.).

Coarse-grained, subophitic rocks of low color index (Du Toit, 1939, pl. XXXV, fig. 2). Pyroxene includes both augite and pigeonite, but orthopyroxene is absent. Broad zonal laths of plagioclase average 1.0–1.6 mm. long. Brown biotite is prominent.

Chief characteristics are the coarse grain, absence of both olivine and orthopyroxene, occurrence of biotite, and low color ratio.

(12) *Metasomatic granophyre*—type locality. Rietkop (31° 52′ S. Lat., 20° 19′ E. Long.).

Medium- to coarse-grained granophyric rocks, produced by transfusion of sediment by dolerite magma (Pl. 8, fig. 4).

Pyroxene builds elongated crystals, usually altered to yellow-green serpentine. Plagioclase forms zonal laths averaging 0.5–1.0 mm. long. The rocks are pale gray in the hand specimen and are characterized by long dark needles of serpentine of branching or criss-cross pattern (Pl. 10, fig. 1). Decomposition is widespread.

Chief characteristics are the low color index, pronounced alteration, and conspicuous prisms of serpentine after clinopyroxene.

Table 2 incorporates the average modes of the various types.

Several variations are recognized, while transition types are also common.

Ophitic dolerites in which columnar, magnesian pigeonite occurs in addition to augite clearly form a gradation between the Perdekloof and Blaauwkrans type. A variation of this type is found in the great Bulthouders Bank sill above Beaufort West. The dolerite is further characterized by large, square augite crystals 1.5–2.0 mm. across, with ophitic margins, and numerous small plagioclase inclusions 0.4–0.6 mm. long.

With increasing plagioclase content, the Blaauwkrans type merges into the Kokstad type. Gradations of this type are most common in the Central Karroo between longitudes 24° to 28°. Other rocks represent locally developed differentiates of the Karroo magma. Thus Scholtz (1936, p. 105) records troctolite at Insizwa. Picrites and picrite-dolerites, formed by differentiation in place, were found at Insizwa-Ingeli (Scholtz, 1936) and along the Elephant's Head dike (Poldervaart, 1944). Intrusions of this type, presumably differentiated before emplacement, form a sheet near Naauwpoort (Walker and Poldervaart, 1942b) and two dikes cutting Kheis beds on the farm Tsebe, in the Northwestern area (Rogers, 1907, p. 88). In the Motiton Reserve, at Takoon, a variolitic dike of picrite-dolerite (Fig. 11) traverses both granite and the Black Reef quartzites (Rogers, 1907, p. 87). A variolitic picrite (Fig. 11) was also encountered in the Elephant's Head dike at Mount Fred, where it forms an outcrop in normal picrites (Poldervaart, 1944). A peculiar, medium-grained, foliated picrite-dolerite is found as a zone in the Insizwa sheet, about 500 feet above its base at Umzimhlava Poort (Du Toit, 1910b, p. 130). The rock (Pl. 8, fig. 5) consists essentially of labradorite, olivine, and augite arranged in bands with the longer axes parallel. Olivine occurs mainly as extremely elongated individuals (elongation ratio 10–15).

At Downes Mountain (Calvinia) the authors (1940, p. 116) encountered a thin granophyre vein, regarded as an extreme differentiate of the dolerite magma. Through the kind co-operation of Dr. J. J. Frankel they examined a granophyric vein from Kranskop (Aliwal North). The rock (Fig. 12 A) is deeply altered and consists essentially of micropegmatite, quartz, and turbid feldspar laths. Augite and a pale greenish yellow serpentine occur in very small quantities, while iron ore is found as rodlike individuals.

In the Glen sheet (Graaff Reinet) the main rock is a sodic differentiate of the Karroo magma (Walker and Poldervaart, 1942b, p. 58). In addition to clino- and ortho-

TABLE 2.—*Average modes of Karroo dolerite types*

Type	Perdekloof	Blaauw-krans	Kokstad	Hangnest	Downes Mountain	Hanover	Marginal tholeiite	Kranskop	Picrite and picrite-dolerite	Dolerite pegmatite	Kentani	Metaso-matic granophyre
No. of Modes	13	28	35	14	6	6	9	5	18	6	4	10
Olivine	3.5	2.8	6.1	—	—	1.7	.1	15.6	44.2	—	—	—
Plagioclase	44.3	42.0	47.5	47.8	43.8	48.7	42.3	31.9	27.1	35	58.7	22
Pyroxene	44.9	42.6	38.8	42.7	42.1	44.1	31.6	44.8	23.7	28	18.6	20
Iron ore	3.9	4.5	2.9	2.9	4.3	2.3	5.6	7.7*	3.4	8	3.6	5
Biot. and amph.	.6	.5	.7	.5	.8	1.3	1.3†	—	1.6	4	13.2	3
Micropeg. and quartz mesostasis	2.8	7.6	4.0	6.1	9.0	1.9	30.1	—	—	25	5.9	50

* Includes about 1 per cent picotite.
† Includes chlorophaeite.

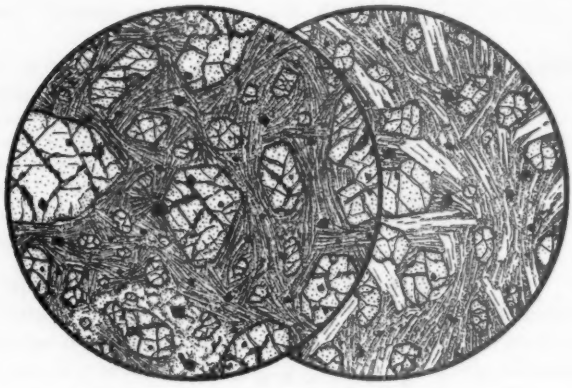


FIGURE 11.—*Variolitic rocks*

A. Variolitic picrite from Mount Fred (Matatiele).
B. Variolitic picrite-dolerite from Takoon (Griqualand West). The texture is caused by the fan-shaped arrangement of interleaved, elongated crystals of plagioclase and augite.

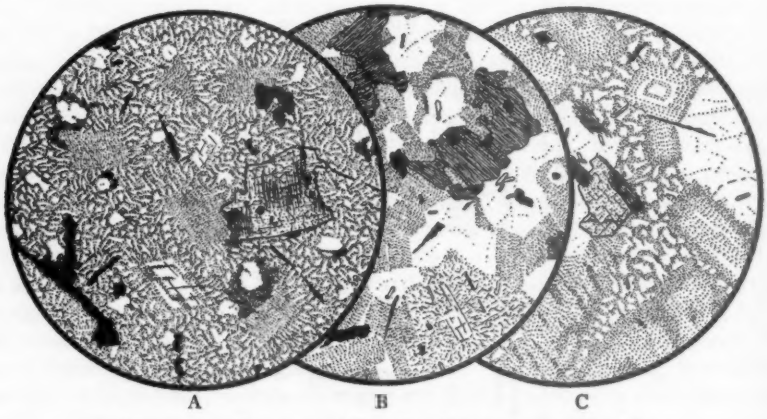


FIGURE 12.—*Karroo granophyres, thought to be of magmatic origin*

A. Vein in dolerite, Poortje (Aliwal North).
B. Vein in dolerite, Hangnest (Calvinia).
C. Vein in dolerite, Cradock.

pyroxene, and plagioclase, this dolerite contains abundant interstitial micropegmatite. The rock has been extensively altered hydrothermally with the development of albite and amphibole.

At New Amalfi, the junior author (1944) encountered a suite of rocks which exhibit differentiation toward iron-rich types (Fig. 13). Thus the coarse-grained dolerites developed here contain ferroaugite, eulite, and ferrohortonolite. As an extreme differentiate, ferroaugite-fayalite granophyre was produced. Here again the rocks show evidence of intense hydrothermal alteration.

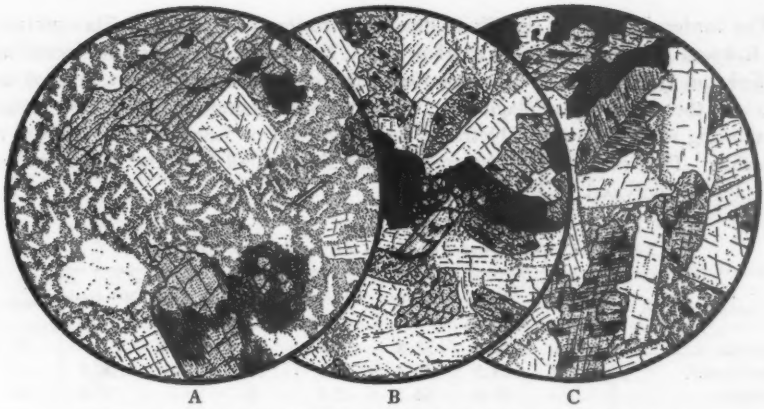


FIGURE 13.—*Iron-rich differentiates of the New Amalfi sheet (Matatiele)*

A. Granophyre with ferroaugite, fayalite, hastingite and brown serpentine.
B. Dolerite with ferroaugite, ferrohortonolite, and eulite.
C. Dolerite with subcalcic ferroaugite and iron-rich pigeonite.

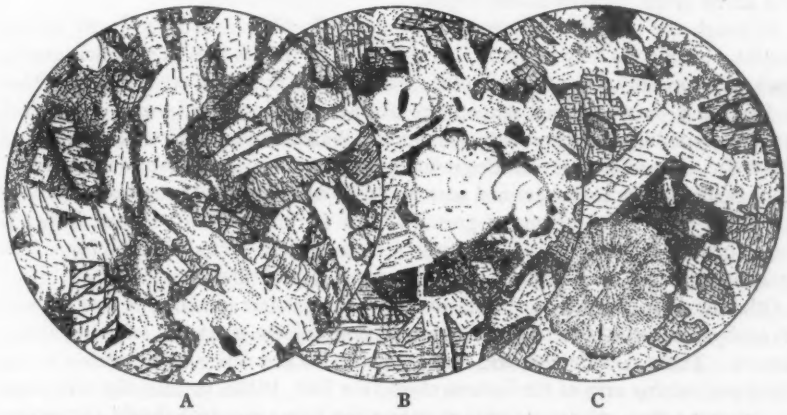


FIGURE 14.—*Tholeiitic modifications of Kokstad type dolerites from Mount Arthur (Kokstad)*

The rocks consist chiefly of plagioclase, pyroxene (augite and columnar pigeonite), and a dense mesostasis. Grains of quartz, marginally crenulated, may occur scattered through the rocks. Near sedimentary xenoliths the tholeiites become vesicular. Alteration is pronounced in all the rocks. (After Poldervaart, 1946).

Some fine-grained olivine dolerites from Natal (*e.g.*, Hulett House, Kearsney) are intergranular (Walker, 1942, p. 217). Augite and pigeonite occur as small, untwinned grains, with a tendency to form polysomatic groups. Plagioclase laths average 0.2–0.4 mm. long.

Kentani-type "dolerites" are of restricted occurrence. Except in the two gap dikes of the Kentani district, these rocks were recognized only in two other localities, viz., on A. 23, near Klipfontein (north of Postmasburg), where a dike cuts the Campbell Rand series (Rogers, 1906, p. 66–67), and in the Queenstown area where they form the core of a dolerite dike (Frankel, personal communication).

The borders of many of the olivine dolerite intrusions of Perdekloof, Blaauwkrans, or Kokstad type are modified to tholeiites. This is most marked in intrusions of Kokstad type, since the conversion to tholeiite is accompanied by a marked decrease in the percentage of plagioclase, without change of grain size (Fig. 14). Table 3, showing average modes, illustrates the conversion of Kokstad type dolerite to tholeiite. The transitional rocks and the tholeiites are all from the Mount Arthur complex (Poldervaart, 1946).

TABLE 3.—*Average modes of Mount Arthur dolerites and tholeiites*

	Olivine	Plagioclase	Pyroxene	Iron ore	Biotite + amph.	Micropeg. + quartz	Mesostasis	No. of modes
Kokstad	6.1	47.5	38.8	2.9	.7	4.0	—	35
Transitional	3.0	43.5	39.4	3.1	.7	—	10.3	9
Tholeiite	—	40.0	36.5	5.7	.8	—	17.0	18

The rocks are characterized by the development of columnar pigeonite and are therefore classified as Blaauwkrans type tholeiites. Olivine, if present, is altered to a bluish-green, pleochroic bowlingite.

Although the pyroxene of metasomatic granophyres is usually replaced by serpentine, the mineral may be fresh. In such cases it proves to be a subcalcic augite, often with cores of columnar pigeonite. Variations in grain size are common. Thus at Rietkop a coarse-grained granophyre overlies a finer-grained variety, in which kernels of undigested Ecca siltstone occur (Walker and Poldervaart, 1942a, p. 286–288). A zone of metasomatic granophyre in the New Amalfi sheet (Poldervaart, 1944) becomes very coarse-grained near vertical joint planes. Along such planes the interlocking of ragged prisms of greenish-black ferromagnesians, several inches in length, may give the rock rude grid-iron structures (Du Toit, 1929a, p. 24). Here again finer-grained varieties form transition zones around siltstone xenoliths.

Other acid rocks which do not show the characteristics of the type metasomatic granophyre as defined here are also found from widely scattered localities in the Karroo. These include the acid veins cutting both dolerite and hornfels in the Nolangeni mining area of the Insizwa sheet (Du Toit, 1910b, p. 131), the acid dikes, several feet across, cutting dolerite sheets on the farms Houdenbek and Cyferwater in the Prieska district (Rogers and Du Toit, 1909a, p. 93–94), and the pipe fillings on the farm Lower Zwart Rand, Carnarvon district (p. 95–96). In thin section the rocks are nearly identical and consist of rounded, corroded quartz areas, clouded feldspar (chiefly orthoclase, but with a little sodic plagioclase), and tufts of chlorite and sericite. Epidote, rutile, sphene, and zircon may occur as accessory minerals. The rocks closely resemble the syntectic material observed by the authors near the terminations of rheomorphic siltstone veins. Scholtz (1936, p. 126–148) mentions several other granite veins and dikes in the Insizwa sheet. Some of these are perhaps of magmatic origin, though they may have been formed by syntexis of Beaufort siltstone.

CHILLED MODIFICATIONS

The basaltic or tachylitic chilled selvages of Karroo dolerite intrusions generally contain a few phenocrysts (Pl. 8, fig. 6; Pl. 9, figs. 1, 2). Volumetric data of representative chilled phases are given in Table 4.

TABLE 4.—*Micrometric analyses of contact basalt phenocrysts*

	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	Average	Average composition of phenocrysts
Olivine	—	2.4	1.3	.1	.5	3.1	.9	1.3	3.1	5.3	1.0	—	—	—	—	1.3	Fa_{15-20}
Plagioclase	10.0	1.3	6.9	1.3	2.7	9.2	.7	3.0	2.5	7.4	4.9	2.0	3.5	3.6	3.4	4.2	An_{70-80}
Augite	—	—	—	.5	.9	1.0	1.2	1.3	1.4	2.5	2.8	.4	—	—	—	.8	$Wo_{39}En_{50}Fs_{11}$
Bronzite	—	—	—	—	—	—	—	—	—	—	—	1.6	.9	1.6	1.8	.4	Of_{15-20}
Total	10.0	3.7	8.2	1.9	4.1	13.3	2.8	5.6	7.0	15.2	8.7	4.0	4.4	5.2	5.2	6.7	

1. Bosrand sill.
2. Theekloof sill.
3. Agterkop, highest sill
4. Lower Hantams sill.
5. Verlaten Kloof sill.
6. Elephant's Head dike.
7. Nooiens Rivier dike.
8. Blaauwkrans sill.
9. Alewyn's Gat sill.
10. Perdekloof sill.
11. Klein Toren sill.
12. Rietkop sill.
13. Hangnest sill.
14. Hangnest dike.
15. Downes Mountain sill.

Chilled phases of dolerites of Blaauwkrans or Kokstad type may show glomeroporphyritic aggregates of plagioclase, which in some instances become prominent (*e.g.*, Bosrand). Variolitic modifications are also encountered, as at Nyembe in the Vryheid district (Campbell Smith, 1912). Frankel (1942) described some younger Karroo dikes, the chilled phases of which contain phenocrysts of olivine, plagioclase, and picotite.

In several instances the basalts show evidence of the passage of late magmatic fluids, presumably through joints and cracks in the chilled selvages. The rocks so affected may be intensely altered, as at New Amalfi, where the phenocrysts of plagioclase, olivine, and augite have been replaced by calcite, serpentine, and talc. At the same time the groundmass has been thoroughly oxidized with the production of innumerable granules of iron ore. At Hangnest the basalts bordering joints show development of poikilitic brown biotite, while the bronzite phenocrysts (Of_{17}) are marked by narrow peripheries richer in $FeSiO_3$ (Of_{40}) or are entirely converted to green pleochroic bastite (Walker and Poldervaart, 1941a, p. 432).

The chilled selvages do not always correspond with the normal dolerite occurring in the same intrusion. Thus, in a sill at Theekloof the chilled phase shows phenocrysts of plagioclase and olivine, while the normal dolerite is a typical bronzite dolerite of the Hangnest type. Normally these dolerites possess chilled phases characterized by phenocrysts of plagioclase and bronzite, but no olivine. The picrite sill at Naauwpoort has chilled selvages of normal olivine basalt, and a few other composite intrusions were encountered.

A remarkable discrepancy is found in the bronzite dolerites of Hangnest and

Downes Mountain (Calvinia district). Apparently the two occurrences belong to the same sill (Rogers and Du Toit, 1903, p. 39), yet the plagioclase of the Downes Mountain rocks is consistently more sodic than that of the Hangnest dolerites and basalts. The chilled basalts of Downes Mountain contain plagioclase phenocrysts of composition An_{66}, while the phenocrysts at Hangnest show an average composition An_{72}. In all other respects the two occurrences are identical.

TABLE 5.—*Modes of feldspar-rich dolerites*

	Average Kokstad type	1	2	3	4	5	6	7
Olivine	6.1	41.0	39.0	11.7	9.2	4.5	3.3	.7
Plagioclase	47.5	39.3	49.0	54.7	54.5	54.5	52.4	55.7
Pyroxene	38.8	16.6	9.6	30.7	33.1	33.6	38.3	36.7
Iron ore	2.9	2.7	1.1	2.5	1.5	2.4	3.2	2.2
Biotite + amph.	.7	.4	1.3	.1	.1	—	1.2	.7
Micropeg. + quartz	4.0	—	—	.3	1.6	5.0	1.6	4.0

1. Picrite, Elephant's Head dike.
2. Troctolite, Insizwa.
3. Execution Rock.
4. Butterworth.
5. Ashton, Mount Arthur Complex.
6. Mahainkwe Poort, Elephant's Head dike.
7. Colesburg.

FELDSPAR-RICH MODIFICATIONS

The authors noted no marked enrichment in feldspar in any intrusion of Karroo age, but slight feldspar enrichment may be found in the lower parts of thick differentiated sheets. Scholtz (1936, p. 100) found a band of troctolite in the Insizwa lopolith which "appears to mark the upper limit of rocks holding more than 25 per cent of olivine". Generally speaking, Kokstad, Hangnest, and Hanover dolerites are richer in plagioclase than other types. The modes of Table 5 illustrate enrichment in plagioclase in Karroo dolerites of the Kokstad type.

ORDER OF CRYSTALLIZATION

Before emplacement the Karroo magma apparently started to precipitate small quantities of either plagioclase, olivine and augite, or plagioclase and bronzite.

After emplacement the magma solidified under hypabyssal, and hence totally different physical conditions. Earlier-formed crystals were partly or wholly resorbed, and crystallization started with the precipitation of small quantities of picotite.

Olivine followed and had a prolonged crystallization period. Generally olivine up to Fa_{60} crystallized before pyroxene, while that of composition Fa_{80-100} separated after pyroxene. All compositions between Fa_{18} and Fa_{100} were recorded in the Karroo, though the study of the New Amalfi sheet revealed a local gap in the crystallization sequence of olivine at approximately Fa_{60-80} (Poldervaart, 1944, p. 94).

Plagioclase then began to crystallize along with olivine, earlier crystals having the composition An_{75-80}. Though one of the first minerals to crystallize, its crystallization period was the longest. The most sodic plagioclase (An_{25-30}) crystallized well after pyroxene. But even this plagioclase grades imperceptibly into the anortho-

clase of the interstitial micropegmatite, thus carrying the crystallization period of the feldspars still further.

Pyroxene started to crystallize soon after the most basic plagioclase. The first crystals were magnesian pigeonite or orthopyroxene. The two minerals follow the trend of olivine, while their crystallization periods overlap that of the clinopyroxenes of high optic axial angle. Magnesian orthopyroxenes (Of_{15-25}) crystallized before augite, intermediate orthopyroxenes (Of_{25-35}) are essentially contemporaneous with augite, while more ferriferous orthopyroxenes crystallized after augite. Orthopyroxenes more ferriferous than Of_{80} were not observed in the Karroo. The most ferriferous orthopyroxenes occur in the upper parts of the New Amalfi sheet, where they crystallized after ferroaugite.

The crystallization of pigeonite was similar. Magnesian pigeonite separated prior to augite, and frequently also before orthopyroxene. Pigeonites of intermediate compositions crystallized simultaneously with augite, while ferropigeonite separated after augite and subcalcic augite, though not after ferroaugite. Pigeonites richer in iron than $Wo_{10}En_{46}Fs_{44}$ were not encountered. Clinopyroxenes of high optic axial angle ($> 30°$) commenced crystallization with the separation of augite, followed by subcalcic augite, and lastly by ferroaugite.

The crystallization of pyroxene was followed by the separation of apatite and titan-magnetite. This in turn was succeeded by the precipitation of brown biotite and amphibole. Later biotites and amphiboles are generally green. Interstitial micropegmatitic intergrowth of quartz and anorthoclase followed. The conversion of olivine to chlorophaeite, brown iddingsite, or green bowlingite, and of bronzite to green bastite, and the albitization of plagioclase presumably belong to the closing stages of crystallization.

SEDIMENTS

The sediments require little description. Most Karroo intrusions are in the Ecca or Beaufort beds. These strata consist either of blue, green, or purple shales and mudstones or of very fine-grained argillaceous sandstones or siltstones. Small calcareous concretions are common, as are cavities, lined with small crystals of quartz and epidote.

The siltstones show angular grains of quartz and turbid plagioclase or potash feldspar, set in an obscure argillaceous matrix containing sericite, biotite, chlorite, serpentine, and irregular grains of magnetite. Minerals found in the heavy residues include zircon, tourmaline, sphene, and colorless or pale-pink garnet.

The upper Beaufort or Burghersdorp sandstones are generally slightly coarser than the lower Beaufort or Ecca siltstones, which have an average grain size of well below 0.1 mm. The arenaceous horizons of the Molteno Beds vary from coarse pebbly grits to medium-grained sandstones. They consist mainly of quartz, microcline, orthoclase, and plagioclase, together with small amounts of iron ore, serpentine, and chlorite. More argillaceous varieties show a gray-brown matrix and varying amounts of iron ore and serpentine. The Red Beds and Cave Sandstones are finer-grained, more argillaceous, and characterized by a ferruginous or calcareous cement.

METAMORPHOSED SEDIMENTS

Torrance (1924) studied the thermal metamorphism of Beaufort mudstones in the Middelburg district, while both Du Toit (1910b, p. 131–136) and Scholtz (1936, p. 122–126) have described the metamorphic aureole of the Insizwa sheet in considerable detail. More recently Mountain (1936; 1943; 1944) and the present authors (1940; 1941a, 1942a) have examined metamorphosed Karroo sediments, incidental to studies of the metasomatism of such sediments by dolerite magma.

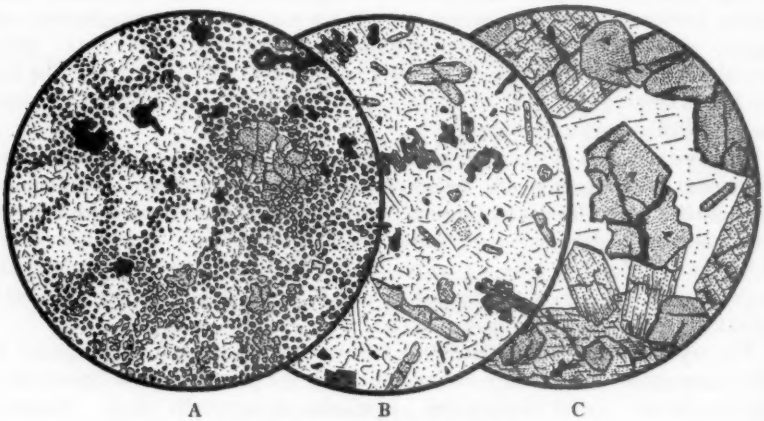


FIGURE 15.—*Hornfelses of the floor of the Insizwa sheet*
Described in detail by du Toit (1910b).
A. Augite-hornfels, Payne's Mining Area.
B. Hypersthene-biotite-hornfels, Nolangeni.
C. Garnet-diopside-hornfels, Mount Ayliff.

The shales and mudstones of the Karroo sediments have been converted by thermal metamorphism to dense black hornstones or lydianite, breaking with conchoidal fracture. More sandy types have been metamorphosed to fine-grained hornfelses, which may contain biotite, magnetite, hypersthene, augite, and cordierite (Fig. 15).

The siltstones and sandstones have been converted to quartzites which show a mosaic of interlocking grains. Upon more extreme metamorphism the alkali feldspar either formed interstitial micropegmatite or porphyroblasts (Scholtz, 1936, p. 124). Thermal metamorphism of coarse feldspathic Molteno grits at Mahainkwe Poort (Matatiele) has produced a rock resembling granite in texture and mineralogy (Du Toit, 1929a, p. 25). The characteristic red color of the Red Beds disappears near dolerite intrusions, owing to the conversion of hematite to magnetite.

In the most extreme cases the sediment may be fused to a clear, light-brown glass. Four such occurrences have been described in the Karroo—*i.e.*, fused Cave Sandstone on the farm Coldbrook near Barkly East (Du Toit, 1904, p. 176); fused Cave Sandstone on the farm Stafelberg's Vlei near Bird's River Siding (Du Toit, 1905a, p. 131); fused Ecca sandstone in the Heilbron district (Young, 1918) (Pl. 9, fig. 3); and fused Molteno sandstone in the Mount Arthur complex (Poldervaart, 1946, p. 94). Fusion of arkose in the Heilbron district is on a very extensive scale. In all examples

the glass contains rounded to angular quartz grains with corrosion borders. The feldspars in the original sandstones appear to have been almost entirely fused to form the glass, together with some of the quartz. Magnetite and zircon are found in all the glasses in small amounts.

Thermal metamorphism of the calcareous pellets in the sediments has given rise to kernels with many interesting and varied assemblages of lime silicates. At Downes Mountain a nodule was found containing large idioblasts of vesuvianite (average diameter 2.2 mm.), set in a fine-grained matrix of calcite, granular grossularite, and wollastonite (Walker and Poldervaart, 1940, p. 166). A calcareous pellet in the Ecca at Rietkop was converted to a kernel of coarsely crystalline quartz and alkali feldspar, with smaller crystals of granular diopside, sphene, calcite, and biotite (Walker and Poldervaart, 1942a, p. 293). Calcareous nodules in the Beaufort beds at Alewyn's Gat are represented by kernels consisting of a core of reddish-brown garnet, surrounded by a zone of fibrous pectolite and turbid thomsonite. This in turn is surrounded by a zone in which aegirine-augite occurs in addition to quartz and alkali feldspar.

According to Scholtz (1936, p. 125), calcareous pellets in the Beaufort beds at Insizwa were converted to wollastonite, grossularite, diopside, alkali feldspar, calcite, hyalite, and almandite. Du Toit (1910b, p. 135–136) mentions several calc-silicate hornfelses, developed in the floor of the Insizwa sheet, and consisting chiefly of garnet, diopside, and wollastonite, with subsidiary zoisite, sphene, chlorite, epidote, and basic plagioclase (Fig. 15). Certain calcareous beds in the Ecca as seen in borehole cores have been mobilized and vesiculated (personal communication by Dr. L. T. Nel).

The metamorphic effects of dolerite intrusions on the coal measures of the Karroo System has been a major tragedy from the economic point of view. By thermal metamorphism the volatile hydrocarbon content of the coals was diminished. Generally bituminous coal was changed to coke, anthracite, or, in rarer cases, to graphite. Blignaut and Furter (1940, p. 67–68) give Rustplaats (Natal) as an example of the the last extreme type of metamorphism. It is of interest to note that the distances through which coals have been anthracitized by the Zuinguin (Perdekloof type) and later Matshongololo (Hangnest type) sheets are six inches and three feet respectively for each foot of dolerite.

ASSIMILATION PRODUCTS

Examples of assimilation of sediment by dolerite magma are decidedly rare in the Karroo. This is the more striking if contrasted with the frequency with which the sediments were mobilized and transfused by the magma.

At Hangnest a xenolith of argillaceous siltstone was assimilated by dolerite of Hangnest type (Walker and Poldervaart, 1941a, p. 447), consisting essentially of plagioclase (An_{63}), augite ($Wo_{38}En_{43}Fs_{19}$), and bronzite (Of_{18}), together with accessory micropegmatite, iron ore, and biotite. The product of assimilation (Fig. 16) is a coarse-grained rock showing broad plagioclase laths (An_{64}) and large prisms of ferrohypersthene (Of_{63}), but no augite. Stumpy crystals of a second-generation plagioclase (An_{48}), with quartz, micropegmatite, and accessory iron ore and biotite are found interstitially. The larger plagioclase laths are crowded with minute rods of

an indeterminable mineral, though their margins are clear. According to MacGregor (1931), this indicates mild contact metamorphism.

At Mount Arthur a pigeonite-rich rock was formed by reaction of Kokstad type dolerite upon the solidified Marginal tholeiites of an earlier intrusion (Poldervaart, 1946, p. 36). The normal Marginal tholeiites consist essentially of plagioclase (An_{64}),

FIGURE 16.—*Assimilation product of argillaceous sediment and bronzite dolerite at Hangnest (Calvinia)*

FIGURE 17.—*Assimilation product of marginal tholeiite and Kokstad type dolerite at Mount Arthur (Kokstad)*
(After Poldervaart, 1946)

subcalcic ferroaugite ($Wo_{33}En_{36}Fs_{31}$), and a dark mesostasis. Iron ore, opaline silica, and chlorophaeite are the accessory minerals. The olivine dolerite magma precipitated olivine (Fa_{28}), plagioclase (An_{72}), and bronzite (Of_{20}) as its earlier constituents. The reaction product (Fig. 17) is a pigeonite tholeiite consisting chiefly of plagioclase (An_{60}), pigeonite ($Wo_{12}En_{59}Fs_{29}$), ferroaugite ($Wo_{31}En_{36}Fs_{33}$), and mesostasis. Pigeonite forms roughly 40 per cent of the total pyroxene.

RHEOMORPHIC VEINS

In conformity with the nature of the Karroo sediments, rheomorphic veins in dolerite fall into two groups—hornfels veins and sandstone or siltstone veins. The former are more common in the Karroo, as the more argillaceous sediments were apparently mobilized with greater ease than the purer sandstones.

Many of the hornfels veins show flow banding, indicating that the material was at least plastic during injection. The chilled selvages of the dolerite intrusions often show an accumulation of fine magnetite dust along the junction with the veins. Away from the margin the dolerite shows no such effects but retains its normal grain size right up to the veins.

The siltstone veins (Pl. 9, fig. 4) are uniform in appearance, and flow banding is absent. Besides the minerals of the normal siltstone, occasional patches of interstitial micropegmatite may be found. Though starting with sharp margins many of the siltstone veins lose their identity after a short distance. Near their terminations,

patches of the mobilized siltstone may occur interstitially within the crystal framework of the dolerite. Many of the siltstone veins are transfused to granophyre, thus losing all character of the original rock. Also, some bodies of metasomatic granophyre become sufficiently mobile to send out veins into the dolerite. Rheomorphic veins of this type were recorded at Alewyn's Gat (Walker and Poldervaart, 1942a, p. 289). One of the granophyre veins rising from below the main body of metasomatic granophyre in this locality contains numerous dark patches which represent remnants of the underlying dolerite pegmatite and the normal dolerite, brecciated by the intrusion of the vein.

An interesting microbreccia at the extreme end of a stoped siltstone block near the upper contact at Hangnest (Pl. 9, fig. 5) consists of angular fragments of basalt embedded in a matrix of mobilized Beaufort siltstone. Apparently the mobilized sediment invaded the dolerite *en masse* and brecciated the chilled selvages. A body of metasomatic granophyre occurs along the same strike line, only a few feet from the breccia (Fig. 10).

MINERALOGY

GENERAL STATEMENT

The mineralogy of the Karroo dolerites, like that of other dolerite provinces, is simple in that the chief constituents are plagioclase, pyroxenes, and olivine. Although the plagioclase and olivine series have been fully studied and present no difficulties, the pyroxenes are complex in their mutual relations. Much remains to be learned concerning the paragenesis of the monoclinic pyroxenes and concerning the relationship of pigeonite toward augite and orthopyroxene. This section is devoted to these problems, to the solution of which the Karroo magmatic province makes a definite contribution.

METHODS

All refractive indices were determined by immersion methods, using sodium light. The indices of the oils used were checked after each determination on a Leitz-Jelley refractometer. The accuracy of the determinations is $\pm$ 0.003. Axial angles and extinction angles were measured on a four-circle universal stage. The probable accuracy is $\pm$ 2°, though the error increases for minerals of high refractive indices. Modal compositions were determined partly with a Dollar integrating stage, partly with a Leitz stage. Except for dolerite pegmatites and metasomatic granophyres, the lengths of the traverses were over 200 times greater than the longest continuous intercept made by any mineral in the sections measured. Hence, the probable error is 1–2 per cent.

A sieved crush was made of all dolerites optically examined. This was separated by bromoform into two fractions, dried, and the iron ore removed with a hand magnet. The light fraction was used for the determination of plagioclase, while the ferromagnesian minerals were determined in the heavy fraction. The main object of the optical work was to determine the approximate composition of the minerals. Hence, rarely were all the optical constants determined for one particular mineral.

For the plagioclase the average β was usually estimated from at least 20 grains for every dolerite. For olivines and orthopyroxenes, 2V and γ were determined. The clinopyroxenes were examined in more detail in selected cases, while checks were applied by determining only 2V and γ' in other instances. The value of γ was obtained by adding 0.004 to the observed value for γ' (Tomita, 1934, p. 47).

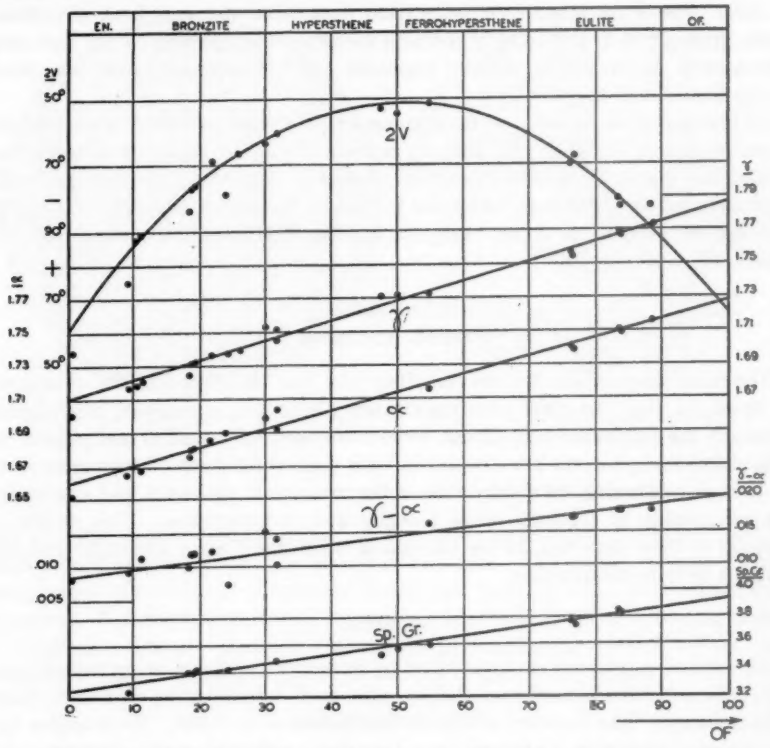


FIGURE 18.—*Optical properties and chemical composition of the orthopyroxene series*
(After Poldervaart, 1947).

The authors agree with Benson (1944, p. 76) that the compositions of all the rock-forming minerals should be expressed in terms of molecular percentages. Although this has been the normal procedure for members of the plagioclase and orthopyroxene series, compositions of olivines and clinopyroxenes are frequently given as weight percentages. In this work all mineral compositions are stated as molecular percentages.

The compositions of olivines were obtained by plotting the γ values on Deer and Wager's diagram (1939, p. 21). The values for 2V were used to check the accuracy of the determinations. A similar procedure was followed to ascertain the compositions of orthopyroxenes. Here the values for γ and 2V were plotted on curves designed by the junior author (1947b) from data given in the literature (Fig. 18). This

procedure was preferred, as discrepancies were noted in the diagrams given by Henry (1935, p. 223) and by Hess and Phillips (1940, p. 280).

The compositions of clinopyroxenes of high axial angles ($2V > 30°$) were determined by the intersections of the γ curves given by Deer and Wager (1938, p. 20) with the course of crystallization as determined by Hess (1941, p. 585). The authors realize this is not a very satisfactory method, but they consider it an improvement

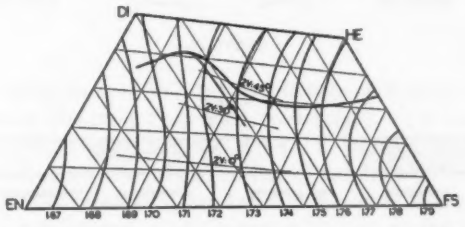


FIGURE 19.—*Course of crystallization and γ curves of nonaluminous, nontitaniferous clinopyroxenes*

on the method formerly adopted—*i.e.*, of plotting the values of γ against those of 2V and obtaining the composition by their intersections (Poldervaart, 1947a). The optic axial angle of clinopyroxenes is strongly affected by even small admixtures of TiO_2 (Kuno, 1936), while Tomita's curves for the values of 2V (1934) were found to be unreliable. The optic axial angles of Karroo clinopyroxenes are consistently lower than those indicated by Tomita's curves. This same tendency is observed in an examination of superior clinopyroxene analyses in the literature. Hess's curve indicating the course of crystallization of clinopryoxenes was slightly modified by the present authors to accommodate the very marked zoning (2V: 48°–30°) observed in some Karroo pyroxenes (Fig. 19). Thus the downward line was added to indicate compositions of clinopyroxenes showing the range of optical constants 2V: 48°–30°, γ: 1.717–1.731. The presence of some such subsidiary line is indicated by the fact that pyroxenes having a γ index of, say, 1.730 were found to have an optic axial angle of either 31° or 44°. It is further indicated by some of the analyses found in the literature. The authors must emphasize, however, that this method gives only very rough approximations of the clinopyroxene compositions and was adopted only in the absence of a better method.

Pigeonites were determined in the orthodox manner—*i.e.*, by plotting the γ values against those of 2V.

NOMENCLATURE

The need for uniformity in nomenclature and subdivision of isomorphous groups of minerals is illustrated by the present confusion in the nomenclature of the pyroxenes, as summarized so well by Benson (1944). Hence, in naming the various members of the different mineral groups, the present writers have tried to preserve uniformity and to follow previous usage of terms.

It appears logical to subdivide the three continuous series of solid solutions—*i.e.*, the plagioclase, olivine, and orthopyroxene series—in a similar manner. This has

already been done for the plagioclase and olivine series (Deer and Wager, 1939, p. 25). The junior author has extended this method of classification to the orthopyroxenes (1947b, Fig. 20). A new name was necessary to cover the range Of_{70-90}.

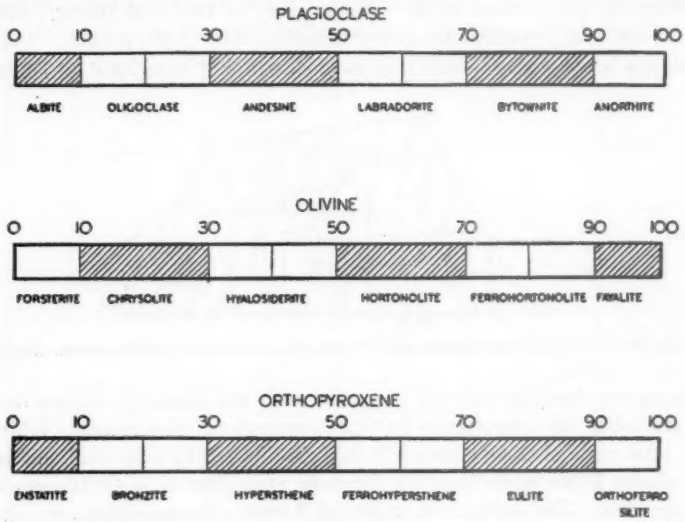


FIGURE 20.—*Nomenclature of the plagioclase, olivine, and orthopyroxene series*
(After Poldervaart, 1947).

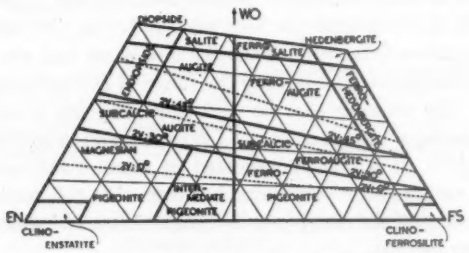


FIGURE 21.—*Nomenclature of the clinopyroxenes*
(After Poldervaart, 1947).

Since analyzed orthopyroxenes of this composition have been described mainly from eulysites, the name "eulite" is here proposed.

In the nomenclature of clinopyroxenes the authors have followed, with but slight modifications (Fig. 21), the classification suggested by Hess (1941, p. 518) and subsequently extended by Benson (1944, p. 111). They suggest the addition of a new subdivision, "ferrohedenbergite", to counterbalance the endiopside field and to indicate the end members of the normal magmatic clinopyroxene series. Though the main divisions are kept, the authors also prefer to separate the salite-intermediate pigeonite fields and the ferrosalite-ferropigeonite fields by a straight line through the

center of the En-Fs line. Benson adopts the lines 2V: 45° and 2V: 30° as the boundaries between (1) the augites and ferroaugites; (2) their subcalcic equivalents; and (3) the pigeonite field.

The position of the lines was taken from Tomita's diagrams. Some alteration will be found necessary as more data become available, but from their observations on the Karroo pyroxenes the authors have drawn in the probable positions of the lines 2V: 45° and 2V: 30° as broken lines.

OLIVINES

No exceptional relations are exhibited by olivines of Karroo dolerites, and all compositions between Fa_{15} and Fa_{100} were encountered. Yet locally a gap may exist in the crystallization sequence of the series. In the New Amalfi sheet a break in the sequence was found between Fa_{50} and Fa_{80} (Poldervaart, 1944, p. 94). Wager and Deer (1939, p. 101) recorded a gap between Fa_{40} and Fa_{60} in the Skaergaard intrusion. Bowen and Schairer (1935) postulated a gap in the crystallization sequence from their investigation of the system MgO-FeO-SiO_2.

Magnesian olivines occur as large, well-developed crystals, while the fayalitic members of the series form small, rounded, or even subophitic crystals. In picrites the terminal faces of the crystals are sometimes welded together, producing elongated crystals with approximately the same optical orientation. They are generally free from inclusions, but in picrites may contain small octahedra of picotite or fine dendritic intergrowths and tabular inclusions of magnetite. The latter are normally parallel to [100] or [001]. Zoning of olivines was observed only in one example from Mount Arthur. Generally the olivines within one thin section show a range in composition, though individual crystals are of uniform composition. In accordance with Tomkeieff's conclusions (1939, p. 243), the range in composition is narrow at the magnesian end, widens for the intermediate olivines, and narrows again toward the fayalite end of the series. Zoning of the olivine in the Mount Arthur dolerite follows the normal trend—*i.e.*, enrichment in fayalite toward the margin of crystals. The greatest variation observed was Fa_{32-42}. The phenomenon is shown by differences in polarization colors and variations in the extinction angle in sections not cut parallel to one of the three principal directions. The larger, strongly zoned crystals are surrounded by numerous minute granules of olivine. None of the other olivine dolerites of the Mount Arthur complex contain zoned olivine. The zoning in the one case is ascribed to a local acceleration in the crystallization of olivine, presumably caused by magmatic reaction with a quartzite xenolith in the immediate vicinity.

Three chief modes of conversion of olivine were recorded in the Karroo:

1. Magmatic stage. By reaction with the magma, olivine may be converted to either orthopyroxene or pigeonite. Whether orthopyroxene or pigeonite is formed appears to depend solely on the temperature of the magma.
2. Hydrothermal stage. Under the influence of a concentration of hyperfusible elements, olivine is frequently altered to green bowlingite or brown iddingsite.
3. Atmospheric stage. Through weathering olivine may be converted to yellow or green, nonpleochroic serpentines and iron oxides.

FELDSPARS

In all normal Karroo dolerites plagioclase is the most abundant mineral by volume. It forms zonal laths, twinned most frequently on the albite law (Table 6).

Generally the more ophitic the pyroxene, the smaller are the average dimensions of the plagioclase laths, the greater the range in size, and the greater the average elon-

TABLE 6.—*Relative frequency of plagioclase twinning laws in per cent*
(Data from Nell and Brink, 1944, p. 37)

Source	No. of determinations	Albite	Carlsbad	Acline-pericline	Roc Tourné	Other twinning laws
Scholtz (1936)	120	49	14	22	11	4
de Swardt and Murray (1944)	124	44	17	15	16	8
van Eeden (1937)	18	22	44	6	28	—
Total	262	45	17	17	15	6

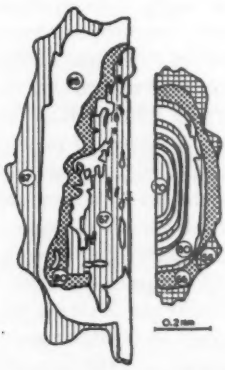


FIGURE 22.—*Zoning in plagioclase of the Insiswa sheet*
(After Scholtz, 1936).

gation of the crystals. Thus in Perdekloof dolerites the average ratio of length to breadth is about 8, whereas this value drops to between 3 and 4 in intrusions of Blaauwkrans, Kokstad, or Hangnest type. Generally the size of the plagioclase laths is seriate, yet sometimes a tendency toward crystallization in two generations is observed (Scholtz, 1936, p. 109).

In picrites plagioclase may occur as slender zonal laths or as large poikilitic crystals of more uniform composition. The laths were formed by crystallization in a magma undergoing movement. The poikilitic habit is ascribed to crystallization in a quiescent magma. Hence picrites containing poikilitic plagioclase were formed *in situ*, while picrites with small plagioclase laths were formed by differentiation prior to emplacement.

Zoning is extremely variable, and examples of resorption and repair, continuous, rhythmic, and reversed zoning may be found within the same sections (Fig. 22)[3].

De Swardt and Murray (1944, p. 6) figure another example of resorption and repair, with core An_{88}, mantle An_{80}, and outer edge again An_{88}. Average compositions of plagioclase in the various types are given in Table 15.

TABLE 7.—*Analysis and optical properties of bronzite from Hanover (C. P.)*

Bronzite from Karroo dolerite at Hanover (C. P.)

Analysis		100 per cent Anh.	Atomic	Ratio	No. mols. on basis 6(O,OH)	
SiO_2	52.86	53.33	$Si^{\cdots}$	.889	1.898	.098, .075 } 1.996
TiO_2	.26	.26	$Al^{\cdots}$	.081	.173	
Al_2O_3	4.08	4.12	$Fe^{\cdots}$	.016	.034	
Fe_2O_3	1.26	1.27	$Fe^{\cdot\cdot}$	.157	.336	
FeO	11.23	11.33	$Mg^{\cdot\cdot}$	.677	1.444	} 1.995
MnO	.23	.23	$Ca^{\cdot\cdot}$	.043	.092	
MgO	26.82	27.06	$Mn^{\cdot\cdot}$	.003	.007	
CaO	2.38	2.40	$Ti^{\cdot\cdot\cdot\cdot}$	.003	.007	
Na_2O	tr.	—	O''	2.810		
K_2O	tr.	—				
P_2O_5	nil	—	*Molecular formula**			
H_2O+	.53	100.00	$Wo_{4.9}$ $En_{78.9}$ $Fs_{16.2}$ } or $Of_{19.1}$			
H_2O-	.07		*Formula*			
Total	99.72		(Mg, Fe···, Fe··, Ca, Mn, Ti, Al) [(Si, Al)$_2$O$_6$]			

Color: Green γ:1.692 $2V_\alpha$: 77°
Pleochroism: None α:1.680 Sp. Gr.: 3.42

* In calculating the composition the molecular proportions of MnO were added to those of FeO, while all other admixtures were neglected.

Plagioclase phenocrysts of chilled phases may occur as isolated crystals or as glomeroporphyritic aggregates. In the latter case the crystals may show intricate zonal phenomena and may include augite subpoikilitically. Remnants of these aggregates are encountered in the finer-grained dolerites.

Interstitial alkali feldspar may occur in the dolerites, normally in a micrographic intergrowth with quartz. It is usually turbid, but in a few cases could be identified as a soda orthoclase of low optic axial angle. In the more acidic rocks the zonal plagioclase laths often merge into surrounding haloes of micropegmatite.

ORTHOPYROXENES

In Table 7 an analysis of magnesian orthopyroxene from the Hanover dolerite is given with its physical constants (Poldervaart, 1947b).

[3] Guimarães (1946) believes that zoning of plagioclase in dolerites is connected with the stability relations of ortho- and clinopyroxene.

The bronzite may show the "graphic intergrowth" along its margins on an extremely fine scale. Many crystals have a peculiar patchy extinction, which is ascribed to the occurrence of the intergrowth on a submicroscopic scale. The rock analysis and mode of the Hanover dolerite are given in Table 15, no. 34.

Apart from the CaO in the Hanover bronzite the analysis is quite normal. Al_2O_3 and Fe_2O_3 are the chief admixtures, the former partly replacing silica. As was shown by Washington and Merwin (1923), both Al_2O_3 and Fe_2O_3 may occur dissolved in the orthopyroxene molecule.

The abnormal optical properties of orthopyroxenes within the range Of_{15-35} has been noted by several authors. There appears to be a gradation in the phenomena observed. In rapidly cooled rocks the mineral generally extinguishes uniformly. In dolerites the orthopyroxene may show a peculiar uneven extinction or, more commonly, sets of finely ruled lamellae and irregular blebs of a mineral of extinction and double refraction different from the orthopyroxene host (Pl. 9, fig. 6).[4] It may occur over the entire crystal, or only in its center, or may be confined to the margins. Lewis (1908) first commented upon the resemblance to the graphic intergrowth of quartz and orthoclase. Subsequently the senior author (1940) adopted the term "graphic intergrowth" to describe the phenomenon. The uneven, patchy extinction represents the "graphic intergrowth" on a submicroscopic scale. In norites the orthopyroxene frequently shows one, two, or three sets of well-developed lamellae, while the irregular blebs are absent. The structure is on a much coarser scale and is much more regular than the "graphic intergrowth". Hence the term "lamellar intergrowth" was introduced to distinguish the phenomenon, as observed in plutonic rocks, from that found in hypabyssal rocks. The above structures are not to be confused with the twinning occasionally observed in orthopyroxenes (*e.g.*, Scholtz, 1936, p. 112–114, Pl. 23, fig. 9). Hess and Phillips (1938) offered two alternative explanations for the development of the "graphic" or the "lamellar" intergrowth. The first explanation is that pigeonite crystallized first, and on slow cooling inverted to orthopyroxene with the segregation of diopside in the form of the intergrowth. The alternative holds that an orthopyroxene separated which was capable of holding in solution $CaSiO_3$ to the extent of about 9 per cent, but unmixed on further cooling. The former relationship was further elucidated in a diagram by Hess (1941, p. 583) (Fig. 23).

The present authors tested this theory for the Karroo orthopyroxenes and can confirm, and in part extend it.

CONVERSION OF PIGEONITE TO ORTHOPYROXENE: As noted, magnesian olivine may react with the magma to form either pigeonite or orthopyroxene. Pigeonite so produced is unstable at lower temperatures and inverts to orthopyroxene upon slow cooling. This orthopyroxene shows the "graphic intergrowth" and may contain rounded remnants of pigeonite (Fig. 24, A and B). Two exceptions are noted: (1) Upon rapid cooling pigeonite is preserved as a metastable phase, (2) in the presence of a high concentration of volatiles (*i.e.*, water), pigeonite is also preserved, even on slow cooling.

[4] *See also* the photographs of Hess and Phillips (1938, p. 452, Figs. 2, 3), Walker (1940, Pl. 2, Fig. 5), Hess (1941, p. 530), and Edwards (1942, p. 589–590).

EXSOLUTION OF $CaSiO_3$ FROM ORTHOPYROXENE: Orthopyroxene (Of_{15-35}) may contain 0–9 per cent $CaSiO_3$. The percentage of $CaSiO_3$ contained in solution appears to depend upon the rate of crystallization rather than the composition of the magma. (*Cf.* Hess and Phillips, 1940, p. 283.) It is thought that orthopyroxenes, crystallized slowly, would contain little or no $CaSiO_3$. If magmatic conditions moved toward the

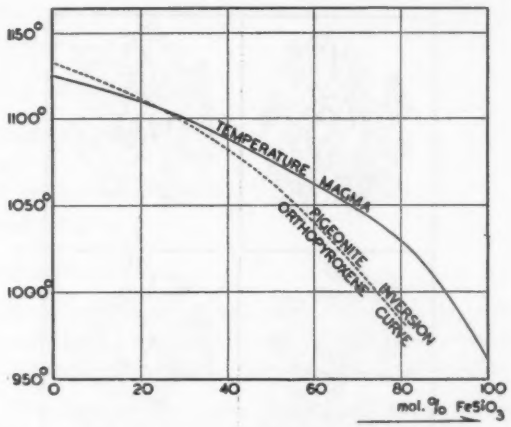


FIGURE 23.—*Pigeonite-orthopyroxene stability relations*
(After Hess, 1941).

stability field of pigeonite, the orthopyroxene would contain more lime in solution upon crystallization. Hence orthopyroxene phenocrysts do not necessarily contain lime.

Clearly, the presence of lime produces considerable strain in the crystal structure. The ionic radius of the calcium cation (1.06Å) is far larger than that of either the magnesian (0.78Å) or the iron cation (0.83Å).

Small amounts of lime dissolved in the orthopyroxene would produce the following strain effects with moderately slow cooling:

(1) Lamellar twinning of the type described by Scholtz (1936, p. 112) and Henry (1942)[5].
(2) Exsolution on further cooling, with the production of, first, patchy extinction and, later, "lamellar" or "graphic intergrowth".

Extremely slow cooling would leave little trace of exsolution of $CaSiO_3$; the lime would go into augite or calcic plagioclase.

In the Karroo, orthopyroxenes were found:

(1) as euhedral phenocrysts in the chilled basalts of Hangnest and Downes Mountain dolerites;
(2) as subophitic plates in picrites, picrite-dolerites, and dolerites of the Kokstad type;

[5] Lamellar twinning of this type may also be produced in the absence of dissolved lime, through disorientation caused by strain.

(3) as subophitic, prismatic crystals in dolerites of the Hangnest type;
(4) as small, stumpy, iron-rich prisms of late formation in differentiated rocks at New Amalfi;
(5) as subophitic plates in rocks produced by assimilation;
(6) as prisms or irregular poikiloblasts in the coarser hornfelses.

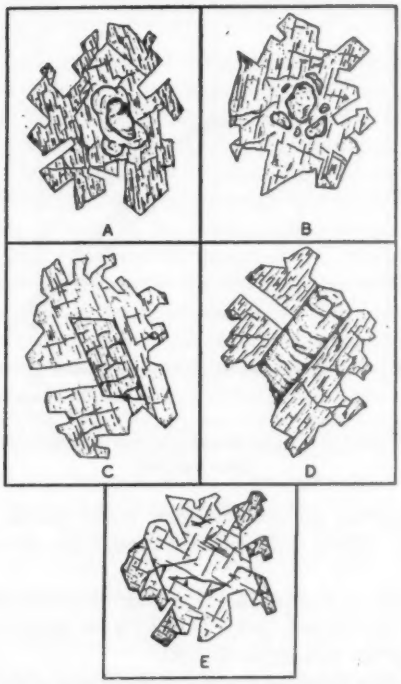


FIGURE 24.—*Relations of pigeonites, orthopyroxenes, and clinopyroxenes of high optic axial angle*

A. Core of olivine, surrounded by successive mantles of pigeonite, bronzite, and augite.
B. Partly resorbed core of pigeonite surrounded by bronzite.
C. Columnar core of pigeonite surrounded by hypersthene.
D. Columnar core of pigeonite surrounded by augite.
E. Core of augite surrounded by ferropigeonite. (After Poldervaart, 1946).

Pleochroism is generally very weak or absent, though one olivine dolerite from Mount Arthur contains moderately pleochroic bronzite, and Scholtz (1936, p. 125) records pelochroic hypersthene from a hornfels at Insizwa.

PIGEONITES

The most magnesian pigeonite encountered in the Karroo was $Wo_{12}En_{66}Fs_{22}$, while he most iron-rich pigeonite was $Wo_4En_{47}Fs_{49}$. The former occurred as columnart cores to subophitic mantles of augite, while the latter formed ophitic mantles to cores of augite (Fig. 24, D and E).

Zoning is common in Karroo pigeonites. The following measurements are examples of the type of zoning:

(1) *Magnesian colorless pigeonite*

core 2V: 12°, optic plane ⊥ (010)
toward margin 2V: 0°
margin 2V: 8°, optic plane ∥ (010)
} average γ: 1.706

Compositions { $Wo_5En_{48}Fs_{26}$ (core) ↓ $Wo_{16}En_{40}Fs_{44}$ (margin)

(2) *Ferriferous brown pigeonite*

core 2V: 0°
margin 2V: 19°, optic plane ⊥ (010)
} average γ: 1.736

Compositions { $Wo_{10}En_{46}Fs_{24}$ (core) ↓ $Wo_4En_{47}Fs_{49}$ (margin)

Magnesian pigeonite frequently forms idiomorphic prisms, often considerably elongated, and sometimes curved. Cleavage is poor, but cross fractures, salite structures, and twinning are common. Late ferriferous pigeonite is usually ophitic

TABLE 8.—*Distribution of orthopyroxene and pigeonite in basaltic and related rocks*

	Plutonic rocks	Picrites and picrite-dolerites	Dolerites of Hangnest type	Phenocrysts of volcanic rocks	Groundmass of volcanic rocks	Dolerites of Kokstad or Hanover type	Dolerites of Perdekloof type	Dolerites of Blauwkrans type	Dolerites of Downes Mountain type	Dolerite pegmatites	Marginal modifications of Kokstad type dolerites	Metasomatic granophyres
Pigeonite.....	—	—	—	—	x	x	x	x	x	x	x	x
Orthopyroxene.....	x	x	x	x	x	x	—	—	—	—	—	—

and crowded with minute granules of iron ore. An intermediate pigeonite (γ:1.714), in the Mount Prospect pegmatite, forms columnar crystals or irregular lamellae in augite. It shows interleaved lamellae and irregular patches of orthopyroxene with graphic intergrowth.

The normal relations of pigeonite and orthopyroxene in different rocks are summarized in Table 8.

However, exceptions may occur. Dolerites of the Hangnest type may contain a little late ferriferous pigeonite. Phenocrysts of pigeonite were found in a Japanese andesite by Kuno (1936, p. 125) and in a semivitreous dacitic inninmorite from Mull by Hallimond (1914). Orthopyroxene lamellae occur in the pigeonite of the Mount Prospect dolerite pegmatite.

Magnesian pigeonite crystallizing prior to orthopyroxene may invert to bronzite or hypersthene upon slow cooling, but the authors believe that volatiles act as catalysts in the preservation of pigeonite, even on slow cooling of the magma. The following supports this theory:

(1) Dolerite pegmatites, metasomatic granophyres, and marginal modifications of Kokstad type dolerites commonly contain pigeonite, while orthopyroxene is not normally found in such rocks. A high concentration of volatiles is generally accepted as essential for the formation of these rocks (Tomkeieff, 1929; Walker and Poldervaart, 1942; Poldervaart, 1946).

(2) In sills of Hangnest type, orthopyroxene occurs in the lower two-thirds of the intrusions, while pigeonite is found in their upper portions, giving rise to the Downes Mountain type of dolerite (Walker and Poldervaart, 1940; 1941a; 1941b). Similar relations are found in the Whin Sill (Holmes and Harwood, 1928), while Edwards (1942, p. 595) notes that "the orthopyroxene of the Tasmanian dolerites is restricted to the chilled margins of the intrusions and to the lower third of the sills." Edwards (1942, p. 589) also finds that in the Mount Sedgwick laccolith the "graphic intergrowth" first appears at the margins of orthopyroxene crystals. At higher levels it occurs toward the interior of the crystals and is enclosed by a margin of clear pyroxene. Identical observations were made by the present authors in the Hangnest sill (1941a, p. 434). Edwards concludes from these relations that "the intergrowth marks an intermediate stage in the passage of hypersthene to plutonic pigeonite."

(3) Intrusions of Kokstad type also contain orthopyroxene in their lower parts, while pigeonite replaces orthopyroxene at higher horizons. The olivine-bearing dolerite of the Palisade sill shows the same relations (Walker, 1940).

(4) Tholeiites of the Mount Arthur complex were evolved from Kokstad type dolerites through the distillation of water from large sedimentary xenoliths. These rocks contain pigeonite to the exclusion of orthopyroxene. All the tholeiites show evidence of pronounced hydrothermal activity (Poldervaart, 1946).

The relations of magnesian orthopyroxene and magnesian pigeonite (Mg: Fe ratio 4:1) under hypabyssal conditions are further summarized in Table 9.[6]

Ferriferous pigeonite is clearly more stable at the temperature prevailing during and after its crystallization than is magnesian pigeonite (Fig. 23).

CLINOPYROXENES OF HIGH OPTIC AXIAL ANGLE

Attempts to separate Karroo pyroxenes for chemical analysis were unsuccessful, except in one case. Normal Karroo dolerites contain two or more pyroxenes, which could not be separated by the usual methods. Moreover, as the pyroxenes are generally strongly zonal, the chemical analysis yields only a relatively valueless average composition. An iron-rich dolerite pegmatite from the Mount Arthur complex contained unzoned subcalcic ferroaugite as the only pyroxene. The chemical analysis of this mineral and its physical constants are given in Table 10. The rock analysis and mode appear in Table 15 (no. 65).

As in most pyroxenes there is some replacement of Si···· by Al···. The analysis is fairly low in sesquioxides, and the composition may be expressed in terms of the wollastonite, enstatite, and ferrosilite molecules. Analyses of composite Karroo pyroxenes indicate that they are all characterized by low sesquioxides.

[6] It is noteworthy that the views here advanced are opposed to Benson's conclusions (1944, p. 109).

TABLE 9.—*Stability relations of orthopyropene and pigeonite*

Temperature of magma (Hess, 1941)		First mineral to crystallize	Condition of magma	Rate of Cooling		
<1110°	Slow rate of crystallization	Normal bronzite	Dry	Slow Fast	Normal Bronzite	
			Wet	Slow	→ Bronzite with graphic intergrowth	→ Pigeonite
				Fast	"Frozen" as normal bronzite or bronzite with graphic intergrowth	
±1110°	High rate of crystallization	Bronzite with dissolved lime (up to 9 per cent $CaSiO_3$)	Dry	Slow	→ Bronzite with graphic intergrowth	→ Normal bronzite
				Fast	"Frozen" as lime bronzite (*unstable*) or as bronzite with graphic intergrowth	
			Wet	Slow	→ Bronzite with graphic intergrowth	→ Pigeonite
				Fast	"Frozen" as lime bronzite (*unstable*) or as bronzite with graphic intergrowth	
>1110°		Pigeonite	Dry	Slow	→ Bronzite with graphic intergrowth	→ Normal bronzite
				Fast	"Frozen" as pigeonite (*unstable*) or as bronzite with graphic intergrowth	
			Wet	Slow Fast	Pigeonite	

The range in composition most frequently noted in Karroo pyroxenes is shown in Table 11, together with the values for 2V and γ. The rocks in which such pyroxenes normally occur are also included.

TABLE 10.—*Analysis and optical properties of subcalcic ferroaugite from Mount Arthur (Kokstad)*

Subcalcic ferroaugite from dolerite pegmatite							Mount Arthur	
Analysis		100 per cent Anh.	Atomic ratio		No. mols. on basis 6(O,OH)		Molecular composition	Physical constants
SiO_2	49.57	49.67	Si····	.828	1.936	.057 } 1.993	Calculated	Color: Brown
TiO_2	.78	.78	Al···	.028	.066		from	$2V\gamma$: 43°
Al_2O_3	1.36	1.36	Fe···	.018	.042	.009	chemical analysis	
Fe_2O_3	1.38	1.38	Fe··	.305	.714			γ_{AC}: 44°
FeO	21.92	21.96	Mg··	.251	.587	1.992	$Wo_{31.6}$ $En_{30.9}$ $Fs_{37.5}$	γ: 1.738
MnO	.45	.45	Ca··	.257	.601			α: 1.712
MgO	10.03	10.05	Ti····	.010	.023			
CaO	14.32	14.35	Mn··	.007	.016		Inferred	Sp. Gr.: 3.47
H_2O+	.44	100.00	O″	2.565			from γ index	
H_2O-	.04						Wo_{30}	
Na_2O, K_2O	tr.	*Formula*					En_{33}	
Total.....	100.29	$(Mg, Fe···, Fe··, Ca, Ti, Mn, Al)_2[(Si, Al)_2O_6]$					Fs_{37}	

TABLE 11.—*Optical properties and chemical composition of common Karroo clinopyroxenes*

	2V	γ	Composition inferred	Rock
←Augite→	51°	1.704	$Wo_{39}En_{50}Fs_{11}$	In picrites, and as phenocrysts in chilled basalts
	50°	1.715	$Wo_{41}En_{43}Fs_{17}$	
	49°	1.717	$Wo_{39}En_{42}Fs_{19}$	In most normal Karroo dolerites
	44°	1.725	$Wo_{33}En_{42}Fs_{25}$	
Sub-calcic→augite	30°	1.731	$Wo_{25}En_{41}Fs_{34}$	

The optical properties of the iron-rich clinopyroxenes are more variable. The two alternative courses of crystallization indicate the occurrence of pyroxenes with the same indices of refraction, yet different optic axial angles. As the pyroxenes become more iron-rich their percentage of titania also tends to increase. Kuno (1936) studied the effect of admixtures of titania on the optic axial angle. Examples of variations observed in Karroo pyroxenes are given in Table 12.

Zoning is usually progressive as evidenced by an increase in the refractive indices and a decrease in the optic axial angle. Exceptions are sometimes noted. Thus a

Kokstad type dolerite from Mount Arthur contains pyroxene which shows abnormal zoning. Actual measurements are given in Table 13.

Zoning is continuous from core to margin. γ varies from 1.704 (core) to 1.713 (margin). The rock contains 18 per cent of olivine, altered to bowlingite. Orthopyroxene is absent. The rock analysis and mode are given in Table 15 (no. 13).

TABLE 12.—*Optical properties and chemical composition of iron-rich Karroo clinopyroxenes*

	2V	γ	Composition inferred	Rock	Composition using Kuno's curves for 2V	Color
Subcalcic augite	30°	1.731	Wo_{25} En_{41} Fs_{34}	In dolerites, dolerite pegmatites, and marginal tholeiites.		Pale brown
	46°	1.725	Wo_{36} En_{39} Fs_{25}			
Subcalcic ferroaugite	44°	1.730	Wo_{32} En_{36} Fs_{32}			
	43°	1.738	Wo_{36} En_{33} Fs_{31}			
	50°	1.734	Wo_{31} En_{35} Fs_{34}			
Titaniferous ferroaugite	52°	1.740	Wo_{30} En_{31} Fs_{39}	Iron-rich differentiates of New Amalfi.	Wo_{35} En_{31} Fs_{34}	Purple brown
	56°	1.758	Wo_{30} En_{19} Fs_{51}		Wo_{33} En_{28} Fs_{39}	Green
					Wo_{42} En_{10} Fs_{48}	Green or purple

TABLE 13.—*Abnormal zoning of pyroxene in Mount Arthur dolerite*

	Core	Toward margin	Margin
A	30°	40°	51°
B	32°	41°	51°
C	37°	42°	50°
D	38°	43°	51°

The normal course of crystallization (Fig. 19) shows a marked downward trend in its central part. During this period the pyroxene is enriched in iron at the cost of lime. As shown by Edwards (1942, p. 598), the introduction of small $Fe^{\cdot\cdot}$ ions in place of the large $Ca^{\cdot\cdot}$ ion causes considerable distortion of the structure. Such replacement would be possible until the $Ca^{\cdot\cdot}$ content of the crystal is reduced to a minimum, beyond which internal stress would render the structure unstable. Further iron enrichment can now be achieved only at the cost of magnesia. Hence the course of crystallization alters in a direction parallel with the En-Fs line.

The $Fe^{\cdot\cdot}$ ion is slightly larger than the $Mg^{\cdot\cdot}$ ion. As more $Fe^{\cdot\cdot}$ is introduced into the structure the dimensions of the links between the silicon chains are slightly increased. Thus a small but increasing proportion of $Ca^{\cdot\cdot}$ ions again enters the crystal trawl as the ferrohedenbergite end of the series is approached.

Clinopyroxene is usually ophitic or subophitic. In dolerite pegmatites and in dolerites near chilled contacts the pyroxene becomes elongated and is sometimes strongly curved. This is particularly noticeable in dolerites of the Downes Mountain type (Pl. 7, fig. 5). In picrites the augite may become poikilitic. Ferriferous varieties are frequently subophitic to idiomorphic.

TABLE 14.—*Trace elements in Karroo rocks (Expressed as parts per million)*

Number in Table 15	61	48	17	5	66	90	102	Average* Gabbro	Skaergaard* Magma	Ionic Radii in Å	Approximate Sensitivity	
SO_3	—	1,200	—	—	—	—	—	—	6,200	0.30	—	SO_3
P_2O_5	1,600	2,400	700	1,400	2,200	900	600	2,800	600	0.33	—	P_2O_5
SiO_2	533,800	485,000	504,200	517,800	526,800	594,400	645,200	482,400	478,300	0.39	—	SiO_2
Al_2O_3	156,400	130,100	165,500	153,400	131,200	157,600	166,600	178,800	186,200	0.57	—	Al_2O_3
Ga_2O_3	*16*	*30*	*30*	*35*	*30*	*20*	*20*	*10*	*33*	0.62	7	Ga_2O_3
Cr_2O_3	*300*	*2,000* (2500)	*600*	*400*	*70*	*200*	*150*	*500*	*900*	0.64	1	Cr_2O_3
TiO_2	11,800	10,300	8,400	9,200	19,700	9,200	7,600	9,700	12,900	0.64	—	TiO_2
V_2O_3	*450*	*500*	*450*	*400*	*250*	*70*	*150*	—	*225*	0.66	7	V_2O_3
Fe_2O_3	2,300	14,300	5,200	10,700	21,400	41,100	22,300	31,600	11,600	0.67	—	Fe_2O_3
Li_2O	*20*	*40*	*10*	*17*	*abs.*	*22*	*17*	*20*	*10*	0.78	6	Li_2O
MgO	56,700	129,100	73,100	64,400	51,300	35,600	17,500	75,000	79,200	0.78	—	MgO
NiO	*10*	*250* (300)	*100*	*85*	*32*	*20*	*20*	*200*	*120*	0.78	1	NiO
CoO	*40*	*90*	*35*	*35*	*40*	*25*	*15*	*100*	*75*	0.82	2	CoO
CuO	*25*	*200*	*200*	*100*	*80*	*12*	*40*	—	—	—	—	CuO
FeO	107,600	91,700	96,100	105,800	106,200	34,400	34,400	59,500	88,700	0.83	—	FeO
ZnO	*400*	*60*	*70*	*125*	*n. o.*	*n. o.*	*n. o.*	—	"	0.83	350	ZnO
Sc_2O_3	*25*	*30*	*15*	*40*	*40*	*15*	*15*	*30*	pr.	0.83	—	Sc_2O_3
ZrO_2	—	—	—	—	*130*	*250*	*250*	—	*25*	0.87	13	ZrO_2
MnO	2,500	1,900	700	2,200	2,300	800	—	1,300	900	0.91	—	MnO
Na_2O	17,200	20,300	15,600	25,900	26,800	24,800	28,200	25,500	25,400	0.98	—	Na_2O
Yt_2O_3	—	—	—	—	*40*	*40*	*40*	*3*	"	1.06	15	Yt_2O_3
CaO	93,600	95,200	123,200	98,800	94,400	56,400	12,400	109,900	105,900	1.06	—	CaO
ThO_2	*160*	*30*	*60*	*80*	*n. o.*	*n. o.*	*n. o.*	—	"	1.10	60	ThO_2
La_2O_3	—	—	—	—	*abs.*	*abs.*	*30*	—	"	1.22	15	La_2O_3
SnO	*450*	*200*	*350*	*450*	*300*	*200*	*350*	*200*	*2,000*	1.27	10	SnO
PbO	—	—	—	—	*abs.*	*10*	*10*	—	"	1.32	30	PbO
K_2O	9,500	4,500	1,900	6,600	8,900	19,000	35,400	8,900	2,000	1.33	—	K_2O
BaO	*400*	*80*	*200*	*200*	*200*	*1,300*	*900*	*70*	*30*	1.43	10	BaO
Rb_2O	*20*	*20*	*20*	*20*	*20*	*200*	*100*	*20*	"	1.49	50	Rb_2O

Roman type—chemical analyses.
Italics—Spectrographic determinations.
Analyst for trace elements in Karroo rocks, R. L. Mitchell.
* Quoted from Wager and Mitchell, 1943, p. 286.

1. Basaltic margin, Rietkop, Hangnest type sill.
48. Olivine basalt, center of dike, Poortje (Frankel, 1942, p. 18).
17. Dolerite, Kokstad type, Execution rock sill.
5. Dolerite, Perdekloof type, Alewyn's Gat, Bulthouders Bank Sill.
66. Dolerite pegmatite, same locality.
90. Metasomatic granophyre, same locality.
102. Siltstone xenolith, same locality.

n.o. = not observed
abs. = absent
— = not recorded
tr = present below sensitivity limit
pr. = present

Simple twinning on (110) is common. When combined with basal partings, as in dolerite pegmatites, herringbone structures are produced.

ACCESSORY MINERALS

Biotite.—Biotite is a common accessory of Karroo dolerites. In picrites it is dark brown, while biotites of normal dolerites, chilled basalts, and acid modifications of the Karroo magma are either golden brown or green. The biotite of the picrites in the Elephant's Head dike is deep red brown. According to Hall (1941, p. 29) the red brown color of the biotites is due to a high titania and low magnesia content; magnesia masks the color. The red brown of the biotite found in the Elephant's Head picrites can hardly be due to this combination of high titania and low magnesia content, as the mineral is derived from magnesian olivine or orthopyroxene. It is thought that the color is here caused by small amounts of chromic oxide.

Golden-brown biotite is the most common in dolerites, though some contain only green biotite (*e.g.*, the Blaauwkrans type dolerite of Paalhuis). Often green biotite is found marginally to hexagonal plates of brown biotite. Hall (1941, p. 30) ascribes the change from brown to green to an increase in the FeO content, accompanied by a decrease in MgO. The following optical properties were ascertained:

	Red brown	*Golden brown*	*Green*
pleochroism	γ: deep red brown α: pale brown yellow	golden brown straw yellow	olive green brown yellow
	$2V_\alpha$: 10°	7°	5°
	γ: 1.642	1.637	1.631

Amphibole.—Brown basaltic hornblende or lamprobolite (Rogers, 1940) is common in Karroo dolerites and dolerite pegmatites. It may be molded on clinopyroxene or associated with interstitial micropegmatite. A blue-green amphibole is often associated with the lamprobolite of dolerite pegmatites and probably grades into more sodic varieties. The optical properties are:

	Brown lamprobolite	*Blue green amphibole*
pleochroism	γ: dark brown α: pale yellow brown	blue green pale yellow green
	$2V_\alpha$: 80°	46°
	$\gamma \wedge c$: 16°	11°
	γ: 1.682	1.713

Dark-green bastite was found as a hydrothermal alteration product of bronzite at Hangnest. Hastingsite was recorded in the iron-rich differentiates of New Amalfi.

Iron ore.—Apart from picotite and chrome-magnetite, occurring in the Elephant's Head picrites and dolerites of Kranskop type, the iron ore of Karroo dolerites is invariably of late formation. The bulk of the ore separated when iron enrichment of the magma had reached a maximum. It is found as skeletal crystals molded on pyroxene or as jagged rods associated with the interstitial material. The ore is nearly

always titaniferous, and TiO_2 increases with the iron content of the rocks. Alteration to leucoxene is often observed in the acid modifications of the Karroo magma.

Apatite.—Chlor-apatite is ubiquitous. The mineral builds slender needles which are associated with the interstitial micropegmatite. In most cases the needles penetrate quartz, indicating that it preceded crystallization of the micropegmatite.

Analcite.—Analcite was found in small quantities, filling interstices of a soda-rich tholeiite from the Mount Arthur bell-jar intrusion.

Chlorophaeite.—Chlorophaeite was recorded only in the Marginal tholeiites of the Mount Arthur complex. Here it occurs as pseudomorphs after olivine, in amounts up to 3 per cent by weight.

Iddingsite and bowlingite.—Pleochroic green or brown serpentine of high double refraction often forms pseudomorphs after olivine. Its formation is ascribed to deuteric or hydrothermal activity.

Prehnite.—Joints in the dolerite are often lined with prehnite, calcite, and subsidiary epidote or hornblende. Good examples are found at Beaufort West dam, Coedmore quarries near Durban, and at Kranskop near Aliwal North. It usually replaces plagioclase and occurs as highly zoned individuals showing typical bow-tie structures. The optical properties, as determined by Frankel (1942, p. 4), are: γ:1.642, β:1.626, α:1.614, sp. gr.:2.84

CHEMICAL DATA

Dr. R. L. Mitchell kindly determined the trace elements in several samples of Karroo dolerites, granophyres, and sediments (Table 14).

Table 15 incorporates 114 chemical analyses, norms, and modes of Karroo dolerites, sediments, and metasomatic granophyres. The geographical distribution of the analyzed dolerites is shown in Figure 25. All but one of the analyses for the present authors were made by Mr. F. Herdsman, to whom special thanks are due for his care and promptness.

Insofar as possible the analyses were arranged according to the classification of the Karroo dolerites adopted in this paper, while the data for each type were arranged in increasing order of silica. As may be seen from the table, 88 Karroo dolerites, 11 metasomatic granophyres and mobilized sediments, and 14 Karroo sediments were chemically analyzed. Of the dolerite analyses, 18 have not been published hitherto, 33 were published by the present authors, 12 from the Insizwa sheet were published by Du Toit, Goodchild, and Scholtz, while other authors published a total of 25 analyses. Several doubtful analyses, published by Cohen (1887), Du Toit (1910b), Torrance (1923), and Van der Merwe (1941), were omitted from Table 15.

Table 16 gives the analyses recalculated to 100 per cent, minus H_2O, CO_2, and minor constituents, arranged in order of increasing $\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100$ ratio. This ratio, first used by Wager and Deer (1939), provides a useful index to the degree of fractionation of rocks derived from basaltic magma and is used extensively in this paper.

PETROLOGY

COMPARISONS WITH OTHER MAGMATIC PROVINCES

The Karroo dolerite magma of South Africa is typically tholeiitic (Kennedy, 1933). It is practically identical in chemical composition with the equivalent Stormberg

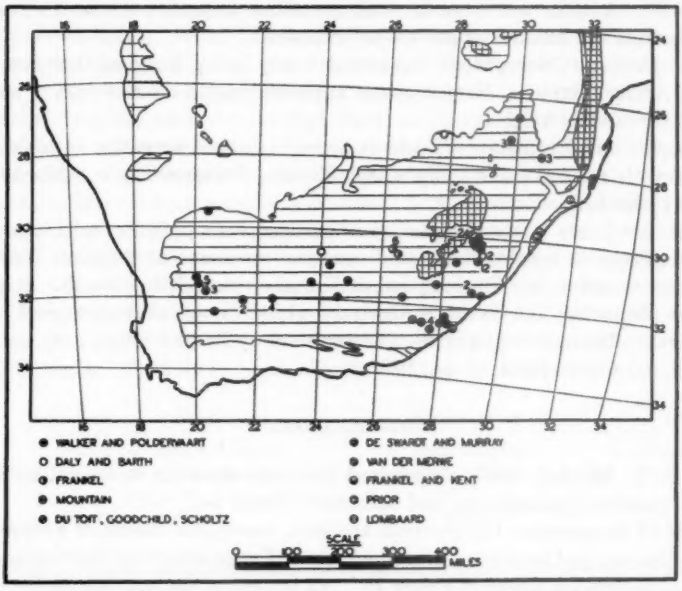


FIGURE 25.—*Distribution of analyzed Karroo dolerites in the Union*

volcanics and shows affinities with many of the other great basalt lava plateaus of the world and with their corresponding intrusive phases. Table 17 gives the average chemical composition of a number of such magmatic provinces. All compositions have been quoted 100 per cent water-free so that satisfactory comparisons may be made; this procedure has been followed in all tables of analyses quoted in this and the following section. This is regarded as essential, for dolerites and basalts show great variations in the amount of combined water which they contain. All obviously differentiated rocks have been excluded from the averages.

The almost exact correspondence of the Karroo dolerites with the Stormberg lavas is to be expected and finds counterparts in other provinces—*e.g.*, the close chemical resemblance of the chilled Palisade diabase to the Watchung traps. In both cases the iron is more oxidized in the effusive phase, where the escape of gas is accelerated; this agrees with the conclusions of Phemister (1934, p. 40–44). Comparisons between the Karroo magma and other doleritic magmas of Trias-Lias age prove interesting. In the Southern Hemisphere the work of Edwards (1942) on the Tasmanian dolerites has provided excellent chemical data which make it clear that the undifferentiated Tasmanian magma (Table 17, no. 7) closely resembles that of the Karroo but is

TABLE 17.—*Average chemical composition of various basaltic provinces*

	Paleozoic		Trias-Lias									Tertiary						Cretaceous				Recent		
	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24
SiO_2	49.8	51.6	51.9	52.5	52.2	50.5	53.3	54.0	51.6	52.2	52.1	51.6	47.4	48.2	52.9	50.8	51.1	50.0	49.5	51.3	49.8	50.6	48.6	50.0
TiO_2	2.0	2.7	1.1	1.0	1.0	3.0	.6	.7	2.0	1.3	1.3	1.6	2.2	1.4	2.2	1.5	1.0	2.9	1.5	1.9	1.7	3.0	3.6	3.1
Al_2O_3	15.6	15.0	15.5	15.4	15.6	14.0	16.4	16.1	13.5	15.4	14.7	14.3	15.6	19.1	15.2	15.3	15.0	14.7	14.1	14.0	15.0	13.9	14.9	13.5
Fe_2O_3	3.5	3.4	1.0	1.2	2.7	3.3	.5	.8	5.3	1.6	3.4	3.5	3.7	1.2	2.4	4.5	3.4	3.4	4.8	3.3	2.7	2.4	4.0	2.7
FeO	8.9	9.6	9.7	9.3	7.9	9.3	8.3	7.4	7.7	8.7	8.6	9.2	9.2	8.7	7.6	6.9	8.4	10.3	10.1	10.1	10.2	11.8	8.1	9.1
MnO	.2	.2	.3	.2	.2	.2	.2	.1	.2	.1	.1	.3	.3	.1	.1	.2	.2	.4	.2	.3	.2	.2	.2	.1
MgO	6.3	4.9	8.2	7.1	7.2	5.4	6.7	7.0	6.3	7.3	7.0	5.3	8.5	7.9	6.7	7.5	6.1	5.5	6.1	5.5	6.5	4.7	6.8	7.8
CaO	9.2	8.9	9.7	10.3	10.1	10.9	11.5	11.1	9.8	10.0	8.9	10.1	10.2	10.7	8.2	8.6	11.1	9.6	9.9	9.8	10.9	8.3	8.3	10.5
Na_2O	2.8	2.4	1.8	2.1	2.0	2.0	1.6	1.8	2.4	2.4	3.0	2.8	2.1	2.4	3.3	3.0	3.0	2.0	2.7	2.8	2.2	3.0	3.6	2.4
K_2O	1.3	1.1	.7	.8	.9	.8	.9	1.0	1.0	.8	.7	1.1	.6	.2	1.0	1.3	.6	1.0	.7	.7	.6	1.3	1.4	.5
P_2O_5	.4	.2	.1	.1	.2	.1	—	—	.2	.2	.2	.3	.2	.1	.4	.4	.1	.2	.4	.3	.2	.8	.5	.3
Number of analyses	10	32	6	43	5	5	6	2	8	4	8	11	7	2	calc.	27	3	4	4	16	17	6	12	23

1. Average Western Province dolerite (Nell and Brink, 1944, p. 56).
2. Average tholeiite and quartz-dolerite of Northern England and Scotland.
3. Average Karroo chilled basalt.
4. Average Karroo dolerite.
5. Average Stormberg lava.
6. Average Victoria Falls basalt (Du Toit, 1939, p. 516).
7. Average Tasmanian chilled basalt (Edwards, 1942, p. 465).
8. Average Antarctica dolerite.
9. Average South American basalt.
10. Average Palisade chilled basalt.
11. Average Watchung basalt.
12. Average Non-Porphyritic Central Magma.
13. Average Hebridean olivine-basalt.
14. Average Skaergaard chilled phase (Wager and Deer, 1939, p. 141).
15. Average northeastern Otago magma (Benson, 1944, p. 104).
16. Average Central Victoria lavas.
17. Average Western Australian tholeiites (Edwards, 1938, p. 7).
18. Average Spitsbergen dolerite (Tyrrell and Sandford, 1933, p. 312).
19. Average Spitsbergen basalt (Tyrrell and Sandford, 1933, p. 312).
20. Average Deccan basalt (Tyrrell and Sandford, 1933, p. 312).
21. Average Siberian Trap.
22. Average Oregon basalt.
23. Average lava from St. Helena, Ascension and Tutuila.
24. Average Hawaii olivine-basalt (Daly, 1944, p. 1392).

slightly more salic and almost identical with certain undifferentiated dolerites of the same age in Antarctica (no. 8). This magmatic type can be matched fairly closely by the analyzed dolerite from the top of Mount Currie (Table 16, no. 33).

The chemical correspondence between the Karroo dolerites and the undifferentiated basic rocks of Trias-Lias age in South America (Table 17, no. 9) is much less close. The latter are considerably richer in Fe_2O_3, total iron, and alkalies, showing in this respect affinities with the late Paleozoic Whin Sill magma of Northern England and Central Scotland (no. 2). On the other hand the Karroo magma is remarkably close in composition to the undifferentiated magma of the Palisades sill, New Jersey (no. 10), and the corresponding effusive phases (no. 11) show an equally marked resemblance. The Rhodesian basalts (no. 6) are distinctly higher in total iron than those of the contemporaneous Stormberg group and have an unusually large content of titania. The Karroo magma is on the whole lower in total iron and alkalies than are later basic magmas of tholeiitic type—*e.g.*, Deccan traps (no. 20), Siberian traps (no. 21), Hebridean N.P.C. magma (no. 12), Spitzbergen (nos. 18, 19), and also the magma of the Western Province dikes (no. 1), of unknown age.

The olivine basalt provinces of the Hebrides, Skaergaard, and certain oceanic islands (nos. 13, 14, 23) have distinctly lower silica than any of the tholeiitic provinces in the table, but the other oxides are very variable and might belong to either of the fundamental magma types. The chilled Skaergaard magma comes closest to that of the Karroo. Magma types of several provinces appear to be intermediate between tholeiitic and olivine basalt—*e.g.*, those of Hawaii (no. 24), Northeast Otago (no. 15), and Central Victoria (no. 16). In spite of the value of Kennedy's conception of the two main world types, it must be recognized that there is an overlap between them. The Karroo magma on the average exhibits definitely tholeiitic characters, yet if the Northern Transkei region had been considered alone the richness in olivine and high MgO of the abundant Kokstad type would have presented an awkward anomaly.

To sum up, the distinctive features of the average Karroo dolerite magma are low total iron and Fe_2O_3, along with fairly high MgO and SiO_2.

DIFFERENTIATION OF THE KARROO MAGMA

Differentiation of the Karroo magma was effected by the following processes, acting singly or jointly: (1) crystal fractionation, (2) gravitational effects, (3) segregation of volatile-rich phases as pegmatitic schlieren, (4) acquisition of resurgent volatiles from wall rocks or xenolithic inclusions, and (5) assimilation or metasomatism of sediments.

In considering crystal fractionation it is necessary to divide the constituent minerals of basaltic rocks into two main groups, viz.: (1) felsic minerals, and (2) ferromagnesian minerals. The plagioclase series forms the major part of group 1, while the pyroxenes constitute the most important part of group 2. Together these two mineral series form about 90 per cent of the average Karroo dolerite, gabbro, or norite.

The thermal relations of the plagioclase series are fully known from experimental data (Bowen, 1913). Lime-soda feldspar separates early in the crystal sequence and has an extensive range. Other minerals of group 1, such as quartz and orthoclase, are of late crystallization. Crystal fractionation, acting on the minerals of group 1,

may therefore be said to result in a gradual increase in SiO_2, Na_2O, and K_2O, combined with a gradual decrease in CaO and Al_2O_3.

Minerals of group 2 include the olivines, orthopyroxenes, pigeonites, clinopyroxenes of high optic axial angle, iron ores, amphiboles, and biotites. Olivines, orthopyroxenes, and pigeonites become progressively enriched in iron as crystallization proceeds. This relationship also holds good for the amphibole and biotite groups, both of which appear late in the crystallization sequence. Matters are complicated by the early crystallization of olivine, generally in excess of the stoichiometric proportion, and by its tendency to settle by gravity and accumulate at lower levels. This is reflected in a steep rise in SiO_2, an equally sharp decrease in MgO, and a more general decrease in FeO.

The order of crystallization of the clinopyroxenes of high optic axial angle has been discussed. It is marked by an increase in FeO at the expense of MgO, as in the other iron-magnesium silicates noted above, except during the middle stage of crystallization, when the increase in FeO is produced at the expense of CaO.

Iron ore appears twice in the crystallization sequence—as early picotite or chrome-magnetite, and as late titano-magnetite or ilmenite. This is reflected in a drop in the proportions of FeO and Fe_2O_3 in the early stages, followed by a prolonged and gradual increase in FeO, Fe_2O_3, and TiO_2 during the middle stage, and terminated by sharp decreases in these constituents during the later stages of crystallization.

To sum up, it may be expected that crystal fractionation, acting on the minerals of group 2, would produce the following effects:

Early stage—a sharp decrease in the proportions of MgO and Fe_2O_3, a more gradual decrease in FeO, and a sharp increase in SiO_2 (crystallization of olivine and chrome-magnetite).

Middle stage—a continued, though more gentle decrease in the proportions of MgO, with gradual increases in FeO, Fe_2O_3, and TiO_2, while SiO_2 shows little variation (crystallization of pyroxenes).

Late stage—sharp declines in the proportions of FeO, Fe_2O_3, and TiO_2, accompanied by a further decrease in MgO and an increase in SiO_2 (crystallization of titano-magnetite, hornblende, and biotite).

The crystallization process and fractionation of basaltic magma is in effect a race between the two mineral groups. Crystal fractionation in group 1 tends to produce a residual magma rich in SiO_2, Na_2O, and K_2O, while the same process, acting on group 2, produces a residuum rich in FeO, Fe_2O_3, and TiO_2. Apparently fractionation in normal basaltic magma affects the minerals of group 2 more than those of group 1, so that iron enrichment is usually more pronounced than enrichment in alkalies. The Karroo magma is no exception to this rule, and iron enrichment dominates, except in the last stages of crystallization, when the reverse is the case (Fig. 27).

Gravitational effects in the magma include the settling of early-formed crystals of olivine, plagioclase, and pyroxene, as well as the upward migration of the acid residuum.

Gravitational settling of olivine was noted in the Insizwa sheet (Scholtz, 1936) and the Elephant's Head dike (Poldervaart, 1944). Some settling of basic labradorite and bytownite also seems to have occurred, both *in situ*, as at Insizwa (Scholtz, 1936,

p. 105), and prior to emplacement, as indicated by the modes produced in Table 5. Appreciable sinking of pyroxene crystals, as noted by Edwards (1942, p. 468) in the Tasmanian dolerites, was not observed in the Karroo, presumably because of the late crystallization of pyroxene, and through crystal interference accentuated by ophitic intergrowths with plagioclase.

The upward migration of the late residuum is more widespread. Its effects were noted by Scholtz (1936, p. 203–207) in the Insizwa sheet, by the present authors at Hangnest (Walker and Poldervaart, 1941a, p. 441) and at New Amalfi (Poldervaart, 1944, p. 98), and is most clearly seen by a comparison of the two analyses from Hangnest (Table 15, nos. 27, 32), taken respectively from 200 and 45 feet below the upper contact.

In a variation diagram showing the differentiation of basalt magma, the effects of crystal fractionation and gravitational concentration should be readily distinguishable. For this purpose the ordinary silica-basic oxide diagram is entirely unsatisfactory, as has been emphasized by Brammall (1933, p. 100), Krokström (1937, p. 271–272), and Wager and Deer (1939, p. 226). Instead, a diagram expressing either iron enrichment (*e.g.*, the ratio $\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100$) or enrichment in alkalies (*e.g.*, the ratio $\frac{K_2O}{Na_2O + K_2O} \times 100$) along the abscissa is far more logical. The two methods of graphical representation supplement one another; the former yields the best results for the early and middle stages of crystallization, while the latter provides the best manner of representing the later stages. The entire crystallization sequence may be represented in a triangular diagram, with variables MgO, FeO, and ($Na_2O + K_2O$). (*Cf.* Wager and Deer, 1939, p. 313.) The limitations of this type of diagram are obvious in the case of basaltic provinces, where a large proportion of the total iron is Fe_2O_3.

Since only a few analyses of "acid" Karroo rocks are available, and some of these are possibly unreliable, only the early and middle stages of crystallization have been represented graphically by plotting the oxide percentages against the ratio $\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100$ (Fig. 26). In addition, the analyses have been plotted on a triangular MgO:FeO:($Na_2O + K_2O$) diagram (Fig. 27). For convenience of reference the crystallization period has been divided into four stages:

(1) $\frac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100 = 25–50.$

Minerals separating:

picotite, chrome-magnetite, and magnesian olivine (Fa_{0-30}), joined later by calcic plagioclase (An_{30-65}).

Main dolerite types:

picrite, picrite-dolerite, Kranskop type dolerite, and basic dolerite of the Kokstad type.

Critical ratio:

45, representing first appearance of plagioclase.

Oxide trends:

SiO_2, Al_2O_3, and CaO show marked increases until 45 and then level out, MgO showing a corresponding decrease. Na_2O increases rapidly until 45 and then more slowly, while K_2O increases steadily

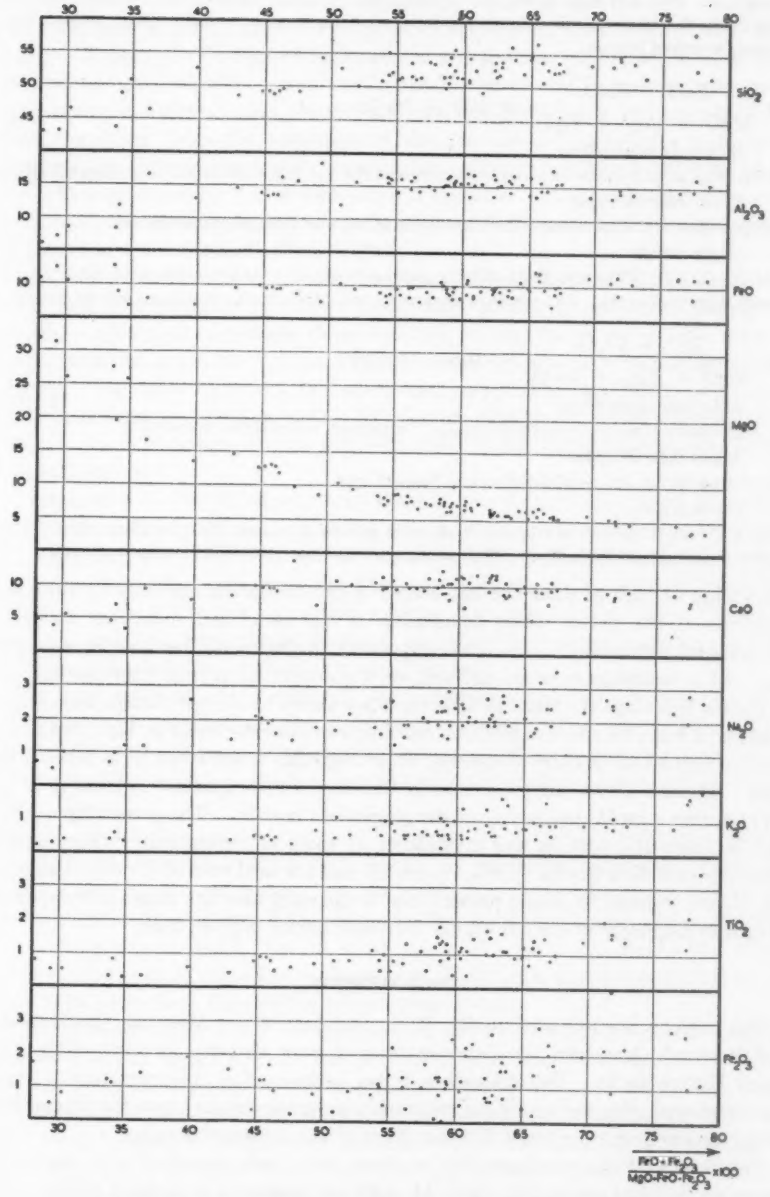


FIGURE 26.—*Chemical variation of the Karroo magma (83 analyses)*

throughout. FeO decreases during the crystallization of early olivine, but increases after 40, while Fe_2O_3, though erratic, shows a gradual increase. TiO_2 shows a gradual decrease, followed by a gradual, sustained increase.

(2) $\dfrac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100 = 50\text{–}70.$

Minerals separating:

Olivine ($Fa_{30\text{-}50}$), plagioclase ($An_{70\text{-}50}$), orthopyroxene ($Of_{18\text{-}25}$), pigeonite, augite, and subcalcic augite.

Main dolerite types:

Kokstad, Hanover, Blaauwkrans, Perdekloof, Hangnest, and Downes Mountain type.

Oxide trends:

SiO_2, Na_2O, K_2O, TiO_2, and Fe_2O_3 show a gradual increase. Al_2O_3 remains level, while CaO increases slightly after 60. FeO reaches a peak at 65, and MgO shows a continued steady decrease.

(3) $\dfrac{FeO + Fe_2O_3}{MgO + FeO + Fe_2O_3} \times 100 = 70\text{–}80.$

Minerals separating:

Iron-rich olivine ($Fa_{80\text{-}90}$), plagioclase ($An_{50\text{-}40}$), ferroaugite, and subcalcic ferroaugite.

Main dolerite types:

Dolerite pegmatite, marginal tholeiite, and Kentani type.

Oxide trends:

SiO_2, K_2O, and TiO_2 show sharp, and Na_2O more gradual increases; Al_2O_3 increases slighlty, while FeO and CaO remain fairly level. MgO continues to decrease, and Fe_2O_3 varies erratically.

(4) Owing to lack of data the last stages of crystallization can only be surmised, and this is the period when transfusion of the associated sediments was most intense and widespread. It is often impossible to distinguish between rheomorphic material of sedimentary origin and acid veins produced by crystal fractionation.

During this stage the trend of differentiation shows an abrupt change; the crystallization of iron-rich silicates and iron ores impoverishes the liquid in FeO, Fe_2O_3, and TiO_2, which all show sharp decreases, while MgO has fallen steadily to below 1 per cent. There is also a sharp decrease in CaO and a more gradual increase in Na_2O, representing a rapid enrichment of the plagioclase in soda. The remaining liquid is rich in silica and alkalies and crystallizes as sodic micropegmatite. The fayalite granophyre of New Amalfi (Table 15, no. 55) and the acid vein of Poortje (Table 15, no. 87) are thought to belong respectively to the early and late stages of this period. A further discussion of the late stages of crystallization is given later.

TRACE ELEMENTS

The authors are indebted to Dr. R. L. Mitchell of the Macaulay Institute for Soil Research, Aberdeen, for determinations of trace elements in seven of the analyzed Karroo rocks. The cathode-layer-arc technique of spectrographic analysis was employed; this method gives reasonably accurate results when the proportions of the trace elements are below 50 times the minimum observable value.

The results of the spectrographic analyses have been combined with the corresponding chemical analyses in Table 14; both are presented in terms of oxides. The order adopted is that of increasing ionic radius.

Four representative dolerites were selected for spectrographic analysis and comprise examples of the Hangnest, Perdekloof-Blaauwkrans, Kokstad, and Kranskop types. Of these, the Perdekloof-Blaauwkrans and Kokstad rocks may be considered as being virtually undifferentiated, while the Hangnest and Kranskop types probably represent differentiation toward the acid and basic poles respectively. The three remaining rocks are associates of the Perdekloof-Blaauwkrans dolerite at Alewyn's Gat, comprising a dolerite pegmatitic schliere, a siltstone xenolith, and a granophyre produced by the metasomatism of the latter. Two comparative analyses are quoted; the first represents average gabbro, and the second the initial Skaergaard magma (Wager and Mitchell, 1943, p. 286).

The major constituents of the undifferentiated Karroo dolerites show these rocks to be slightly more acid than either the average gabbro or the initial Skaergaard magma. Agreement is perhaps closer with the former, except for the low state of oxidation of the iron in the Karroo, a feature which is also notable in the Skaergaard magma. The trace elements, too, show reasonably close agreement in all four analyses, except for SrO and BaO. The former constituent has remarkably high values in the Skaergaard magma, while the latter is distinctly above normal in the Karroo dolerites.

Compared with the undifferentiated dolerite the slightly more basic Kranskop rock is distinctly higher in Cr_2O_3, NiO, and CaO (all of which are associated with the earliest crystal phases) as well as higher in Li_2O, probably in the later ferromagnesian minerals. SrO, BaO, and Rb_2O are all lower, being associated with late feldspars in which the rock is relatively poor. The agreement between the values of Cr_2O_3 and NiO respectively obtained by chemical analysis (Frankel) and by spectrographic analysis (Mitchell) is very satisfactory.

The trace elements of the relatively felsic Hangnest type sill are more or less in accordance with expectations; Cr_2O_3 and NiO are lower than in the undifferentiated types, while BaO is higher. SrO and CaO are about the same, but there is a distinct rise in the value of ThO_2 in the Rietkop rock which is again to be expected in an acid differentiate.

It was hoped that the trace-element values in Alewyn's Gat rocks might throw some light on the origin of the dolerite pegmatite and metasomatic granophyre, but they are only faintly suggestive. As compared with the undifferentiated dolerites, the pegmatitic phase shows little difference, but lower Cr_2O_3 and NiO correspond with the poverty in MgO. On the other hand, SrO, BaO, and Rb_2O which might be expected to rise in value due to abundance of felsic and late ferromagnesian minerals are actually much the same as in the normal dolerite.

The Beaufort siltstone is a feldspathic rock and in comparison with the normal dolerite is much higher in BaO and Rb_2O, but lower in Cr_2O_3 and NiO. These characteristics are repeated in the metasomatic granophyre, but the concentration of BaO and Rb_2O is accentuated in spite of the lower K_2O, a constituent with which they generally show a sympathetic variation. The evidence here is confusing, for in the Skaergaard intrusion, which forms the closest available analogy to the Karroo dolerite suite, Wager and Mitchell (1943, p. 294) record a marked sympathetic variation between $K^{\cdot}$, $Ba^{\cdot\cdot}$, and $Rb^{\cdot}$.

A tentative suggestion is made by the authors that the Hangnest type of dolerite might have arisen by a combination of fractionation and assimilation of siltstone. The high BaO value of the Hangnest type (Rietkop) can scarcely be attributed to simple fractionation on such a slightly differentiated rock, and some slight support is lent to the hypothesis that siltstone rich in that constituent may have been assimilated at depth.

While the values for the trace elements in the suite under consideration are, on the whole, in support of crystal fractionation as a prime factor in the evolution of the dolerite types, the evidence is extremely scanty. Further work in this field would almost certainly prove fruitful.

IRON ENRICHMENT IN BASALT MAGMA

The problem of iron enrichment in basaltic magma has engaged the attention of petrologists for over two decades and has been summed up by Fenner (1931, p. 549):

"... in the competition in a crystallizing magma between the two principal solid solution series—plagioclase feldspars and iron-bearing metasilicates (e.g., pyroxenes)—it is not possible to tell with our present knowledge which side will prevail in the diminishing residual liquid".

Bowen (1928, p. 80–83) favored alkali and silica enrichment, with the production of dioritic and granitic partial magmas, while Fenner has been a staunch protagonist of iron enrichment (1938, p. 367–400). The senior author (1935, p. 151–159), by separation of the fresh residual glass of a chlorophaeite tholeiite, showed that after the crystallization of iron-rich silicates the residue was rich in silica and alkalies, but poor in iron and magnesia. This observation was confirmed quantitatively in a much more satisfactory fashion by Wager and Deer (1939). They showed, in the strongly fractionated Skaergaard magma, that absolute enrichment in iron continued through 95 per cent of the crystallization, but that during the crystallization of the remaining 5 per cent the differentiation trend changed abruptly, producing a residue rich in silica and alkalies, but poor in iron. They do not, however, exclude the possibility of a limited amount of contamination by acid gneiss.

Wager and Deer argued that the trend of the Skaergaard differentiation (*i.e.*, enrichment in iron) is typical of the fractional crystallization of average basalt magma and in a series of ingenious diagrams compared their calculated Skaergaard liquids with analyses of lavas and hypabyssal rocks from Mull and Ardnamurchan (Richie, Thomas, *et al.*, 1930), Daly's averages for the calc-alkaline series, the Breven and Hällefors dikes (Krokström, 1932; 1936), the typical differentiation trend of Central Victoria (Edwards, 1938), the Palisade sill (Lewis, 1908), the Lugar sill (Tyrrell, 1916), the Shiant sills (Walker, 1930), the Glen More ring-dike (Koomans and Kuenen, 1938), and the Carsphairn calc-alkaline hybrid suite (Deer, 1935).

Edwards (1942, p. 605) also found that the

"... trend toward the enrichment of the residual magma in iron relative to magnesia is strongly marked in the Tasmanian dolerites, and ... is so characteristic of basalts in general, that it may be regarded as the normal trend in the differentiation of basaltic magmas".

As has been pointed out, the differentiation trend of the Karroo magma shows the same two features exhibited by that of the Skaergaard magma, viz., absolute

enrichment in iron for the major part of the crystallization period, followed in the last stages by enrichment in silica and alkalies. The changes observed are clearly shown on a triangular diagram with variables MgO, FeO, and ($Na_2O + K_2O$). Thus

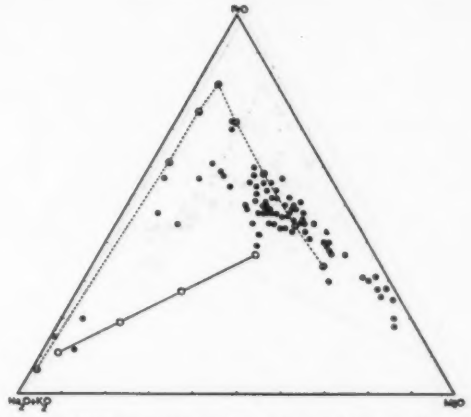


FIGURE 27.—*Differentiation trend of Karroo dolerites (88 analyses)*

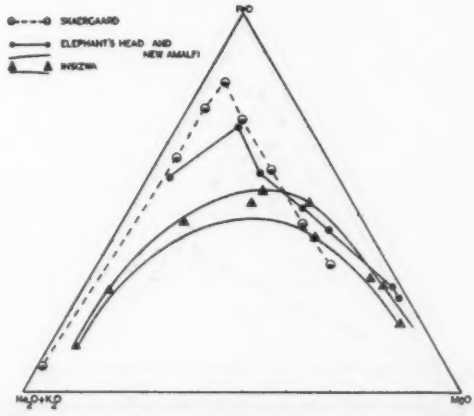


FIGURE 28.— *Comparison of the trend of differentiation of the Skaergaard, the Elephant's Head dike and New Amalfi sheet, and the Insizwa intrusion*

the first part of the trend of differentiation is represented by a line approximately parallel to the MgO—FeO side, while the second part is shown by a line approximately parallel to the FeO—($Na_2O + K_2O$) side. A series of such diagrams has been constructed with the object of extending the comparisons made by Wager and Deer (1939, p. 313–320) to other magmatic provinces. The first of these (Fig. 27) shows all analyses of Karroo picrites, dolerites, dolerite pegmatites, and granophyres plotted on this basis, while Daly's averages of the calc-alkaline series, and Wager

and Deer's liquid line of descent of the Skaergaard magma are included for comparison. In Figure 28 the analyses of Insizwa rocks (Scholtz, 1936) are compared with those of the Elephant's Head dike and New Amalfi sheet (Poldervaart, 1944) and

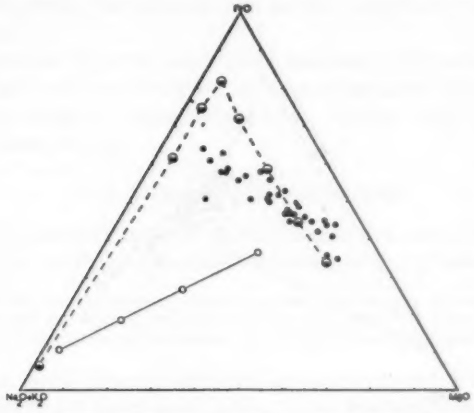


FIGURE 29.—*Differentiation trend of Tasmanian dolerites (34 analyses)*

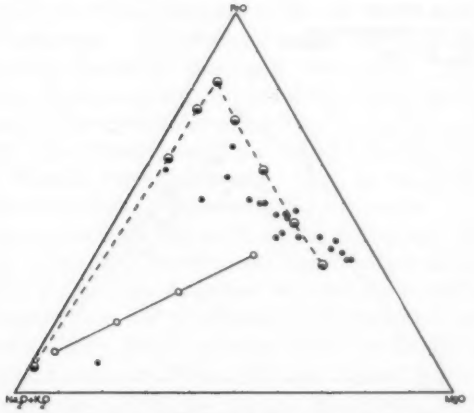


FIGURE 30.—*Differentiation trend of the Palisade sill (21 analyses)*

with the Skaergaard line of descent. It appears that fractionation of the ferromagnesian minerals was strongest in the Skaergaard intrusion, weakest in the Insizwa sheet. The reverse relation holds good for the fractionation of the felsic minerals, in effect the plagioclase series.

Figures 29–33 are comparisons of other magmatic provinces with both the Skaergaard trend of differentiation and Daly's averages of the calc-alkaline series. The analyses represented are Tasmanian dolerites (Edwards, 1942), Palisade sill (Walker, 1940), Siberian traps (Anikeev and Moor, 1939; Levinson-Lessing, 1932;

1935; Moor, 1939; Unksov, 1934)[7], the late Paleozoic quartz dolerites and tholeiites of Scotland and Northern England (Day, 1928; Holmes and Harwood, 1928; 1929; Tomkeieff, 1929; Smythe, 1930; Walker, 1935), and the Hawaiian lavas (Washington,

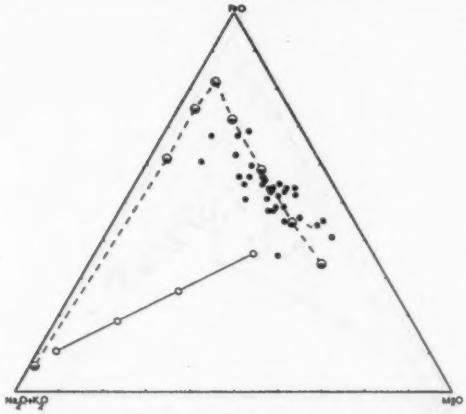


FIGURE 31.—*Differentiation trend of Siberian traps (36 analyses)*

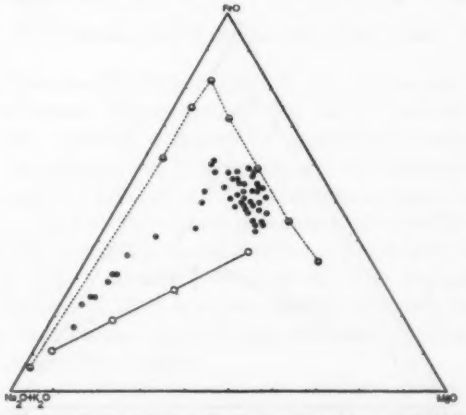


FIGURE 32.—*Differentiation trend of late Paleozoic quartz dolerites and tholeiites of Scotland and northern England (55 analyses)*

1923a–d; Washington and Keyes, 1926; 1928). Iron enrichment is the principal trend of differentiation in all these examples. Thus, during the main period of crystallization, fractionation was more pronounced in the ferromagnesians than in the plagioclase series, and only during the last part of crystallization was there strong fractionation in the feldspars. If the present evidence is regarded in conjunction with that of Wager and Deer (1939, p. 298–324), it may be regarded as established that iron enrichment is the normal trend of differentiation during the greater part of the crystallization of basaltic magma. The differentiation trend

[7] Summaries of these papers were kindly made available to the authors by Dr. Tomkeieff and Dr. Davidson.

toward ($Na_2O + K_2O$) of the Whin sill magma and of the Hawaiian lavas (Figs. 32, 33) is less steep than that of the other provinces. In this respect they represent a transition toward the calc-alkaline line of descent discussed below. The apparently

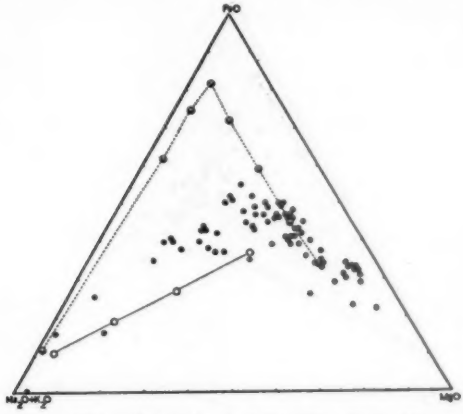


FIGURE 33.—*Differentiation trend of Hawaiian lavas (96 analyses)*

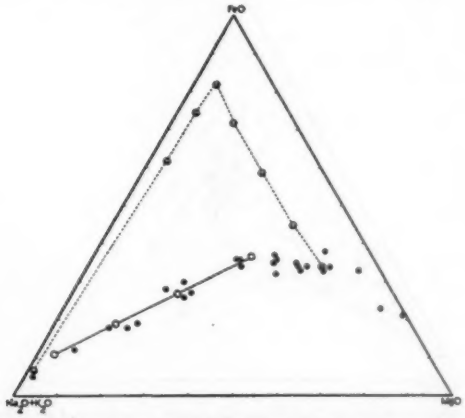


FIGURE 34.—*Differentiation trend of the Garabal Hill—Glen Fyne complex (29 analyses)*

moderate degree of iron enrichment in the Whin sill magma is partly due to the relatively high Fe_2O_3:FeO ratio.

The line of descent of the calc-alkaline suite, as represented by Daly's average basalt, andesite, dacite, and rhyolite, is not in accordance with this trend of differentiation. Wager and Deer (1939, p. 335) believe that "the calc-alkaline series of igneous rocks is, in the main, the result of the mixing of basic and acid material." Edwards (1942, p. 609), likewise considers that the great bulk of calc-alkaline rocks ". . .can have arisen . . . only by the assimilation by tholeiitic magma of large quan-

tities of sialic material, coupled with crystallization-differentiation in subjacent chambers." He considers it possible, however, that the calc-alkaline suite may be produced by fractionation of basalt magma in subjacent chambers, through sinking of iron metasilicates and their displacement by undifferentiated magma from lower levels.

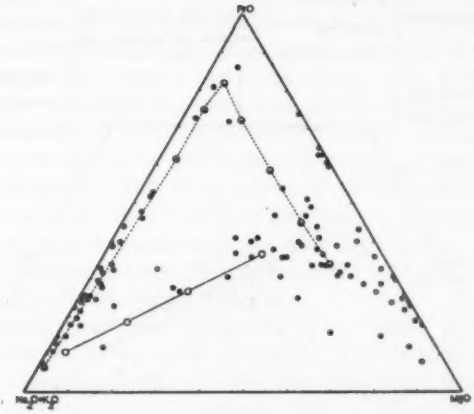


FIGURE 35.—*Differentiation trend of the Bushveld igneous complex (98 analyses)*

The present authors consider it likely that the calc-alkaline suite may be produced by fractionation of basalt magma, though they do not exclude the possibility of assimilation of sialic material, followed by crystallization-differentiation of the hybrid magma. As pointed out, iron enrichment in crystallizing basalt magma means that fractionation affected the ferromagnesians more than the feldspars. Zoning in the pyroxenes is therefore more pronounced than in the plagioclases. The reverse may, however, take place in suites richer in Fe_2O_3 and poorer in FeO, with the production of typical calc-alkaline variation. The Garabal Hill-Glen Fyne igneous complex (Nockolds, 1941) is a case in point (Fig. 34), and Nockolds holds with good reason that normal crystallization-differentiation has been responsible for the bulk of the rocks of the complex.

The Bushveld igneous complex forms a most striking example of a province, in which both types of differentiation are seen (Fig. 35, data from Hall, 1938; Nel, 1940). Rhythmic layering has produced feldspar-rich and ferromagnesian-rich bands, and fractionation, acting on both mineral groups, has resulted in two distinct differentiation trends: (1) a trend similar to that of the Skaergaard magma, and (2) the calc-alkaline line of descent. That iron enrichment is pronounced in the Bushveld lopolith is proved by the occurrence of fayalite-hedenbergite granophyre in the Upper Zone at Tauteshoogte(Boshoff, unpublished paper).

Finally, it may be concluded that fractionation of plateau basalt magma and its intrusive representative generally results in iron enrichment during the main period of crystallization, changing toward enrichment in alkalies during the last stages.

Stronger fractionation of the feldspars than of the ferromagnesian minerals, producing alkali enrichment throughout the crystallization period, is more characteristic of differentiation in calc-alkaline provinces or in basic volcanic associations (Kennedy and Anderson, 1938) with high Fe_2O_3: FeO ratios.

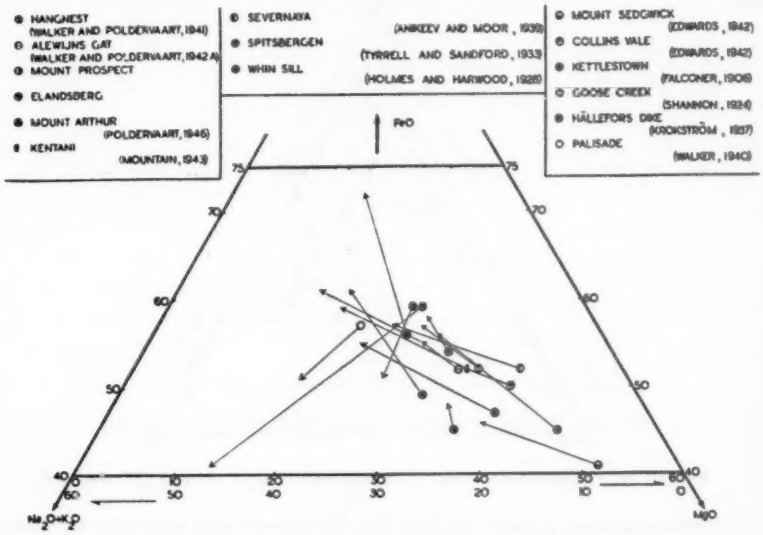


FIGURE 36.—*Chemical variation of dolerites and dolerite pegmatites*

DOLERITE PEGMATITES

In nearly every dolerite suite there are patches and schlieren of relatively coarse rock to which the name "dolerite pegmatite" has been applied. They are found mainly in the upper portions of sills or sheets and do not, in general, exceed a few feet in width. Their form is somewhat irregular, but many of them are lenticular with marked flattening parallel to the contact. Such schlieren and patches of dolerite pegmatite are fairly common in the thicker Karroo sills and are especially conspicuous on freshly blasted exposures, but they are no more abundant than in other dolerite provinces of tholeiitic magma type. Dolerite pegmatite has been recorded in both bronzite and olivine dolerites.

Tomkeieff (1929), when dealing with certain coarse bands and schlieren in the Whin sill, attributed the formation of dolerite pegmatite to the segregation of wet fractions in the magma, the relatively high proportion of volatiles reducing viscosity and thereby promoting the growth of large crystals. Subsequent work has supported this hypothesis, and the present authors believe the dolerite pegmatites of the Karroo originated in this way. Tomkeieff (1929, p. 117) regarded the dolerite pegmatite of the Whin sill as a heteromorphic variety of the normal rock, but it is shown below that, where chemical data are available, the pegmatitic phase belongs to a later stage in the differentiation series than that reached by the parent rock. A triangular diagram (Fig. 36) with MgO, FeO, and (Na_2O + K_2O) as variables has been pre-

pared, on which all analyses of dolerite pegmatite from tholeiitic provinces have been plotted together with those of the normal rock from which they were derived. In each case the pair of cognate points are joined by a straight line with an arrowhead at the dolerite pegmatite end. Each arrow is directed along a path corresponding with the normal differentiation trend of tholeiitic magma—first upward away from the MgO corner toward the FeO apex, and then downward toward the (Na_2O + K_2O) corner. In each case the dolerite pegmatite is further along this path than is the parent rock. The length of the arrow gives a measure of the differentiation producing the dolerite pegmatite. The junior author (1944, p. 106) has suggested that in the absence of dolerite pegmatite schlieren the normal rock becomes more highly fractionated, with the formation of iron-rich types.

The mineralogy and texture of the Karroo dolerite pegmatites is in keeping with their hydrous paragenesis and more advanced differentiation. The elongated pyroxene prisms are often curved and may form graphic intergrowths with plagioclase, while pigeonite occurs to the exclusion of orthopyroxene. The plagioclase is more sodic, and the pyroxene richer in iron than in the normal dolerite, while amphiboles and micropegmatite are relatively abundant.

All the dolerite pegmatites have a higher Fe_2O_3:FeO ratio than their parent rocks. If cooling was slow and if at least 60–80 per cent of the main mass was solid when the pegmatite phase crystallized, this relatively high proportion of ferric oxide agrees with the conclusions reached by Phemister (1934, p. 41) on theoretical grounds. The oxidized state of the iron in the dolerite pegmatite prevented much of it from forming pyroxene and thereby led to the crystallization of considerable amounts of quartz and iron ore.

In at least one case in the Karroo (Walker and Poldervaart, 1942a, p. 289) dolerite pegmatite occurs adjacent to a large xenolithic mass of siltstone. It is extremely doubtful, nevertheless, if the sediment contributed to its formation beyond supplying volatiles which would promote coarse crystallization. *See*, however, in this connection, Alling (1936, p. 247–248) and Wager and Deer (1939, p. 190). The pegmatitic patches described by Phemister (1934, p. 99–119) and others from certain Canadian diabases are scarcely comparable, being associated with purely quartzose sediments of negligible volatile content. All the dolerite pegmatites noted by the present authors are characteristically fresher than any acid syntectic modifications so far described from the Karroo province.

At Downes Mountain, the present authors (1940, p. 165–166) described a thin vein of dolerite pegmatite which passes into granophyre and is situated near the center of a 200-foot sill of Hangnest type dolerite. They regard this as a product of normal crystal fractionation carried a stage further than usual. Similar schlieren have been recorded by the senior author (1940, p. 1075) from the upper part of the Palisade diabase sill and doubtless originated in the same way. Mountain (1943, p. 67) considers that the Downes Mountain example was probably due to the incorporation of sediment, but it is far fresher than any syntectic dolerite yet described from the Karroo province and, furthermore, strongly resembles the differentiates of other tholeiitic magmas which show no tendency to react with the associated sediments.

Although the dolerite pegmatites are much less altered than any of the metasomatic or syntectic rocks, some of them have undergone hydrothermal alteration—mainly albitization. Pectolite and prehnite are fairly common in them, and their crystallization probably belongs to this stage.

LATER STAGES OF CRYSTALLIZATION

The later stages of differentiation in any basaltic province are always more difficult to investigate than the earlier. Abrupt changes in the composition of the residual liquid take place then, as one constituent after another is eliminated, and the solid phases may be altered by hydrothermal fluids. In any case the last fraction will generally take the form of residual injection veins which are often sparse and have frequently suffered decomposition, quite apart from hydrothermal reactions. In the Karroo province a further difficulty is encountered. Even relatively undifferentiated dolerite has a marked tendency to react with the associated sediments, and in the later differentiates, particularly those which have reached or have just passed the stage of maximum iron enrichment, this tendency is greatly accentuated. In consequence, most of the later acid fractions of the Karroo dolerite magma may be shown to have been contaminated through reaction with sediments, and nearly all of the remainder are suspect. In the earlier dolerite pegmatite stage the freshness of the rock forms some criterion as to whether it has been contaminated, but the rocks representing later fractions generally decompose readily, even if they are not contaminated. Thus analyses of later acid veins may not represent accurately the composition of the partial magmas from which they have been formed. Lastly there is the undoubted risk of confusing rheomorphic veins of mobilized sediment with those of igneous origin.

The dolerite pegmatite-granophyre vein of the Downes Mountain sill is one of the very few whose igneous origin is reasonably certain. It is of particular value as providing a transition from the dolerite pegmatite stage to the final quartzo-feldspathic residue. All the minerals are reasonably fresh, and since it follows a sinuous course in the center of the sill and is not associated with xenolithic blocks rheomorphism or contamination seem improbable. Unfortunately, insufficient material was available for chemical analysis. A very thin vein of similar appearance was observed in the upper part of the Hangnest sill (Fig. 12B).

The veins of microgranite cutting olivine gabbro and hornfels in the Insizwa sheet (Du Toit, 1910b, p. 131; Goodchild, 1916; Scholtz, 1936) may represent highly differentiated fractions of the magma, but a rheomorphic or syntectic origin is possible. They resemble acid dikes and pipe fillings from the Carnarvon district, considered below.

Acid veins have also been recorded in the Poortje dike by Dr. J. J. Frankel (personal communication), who very kindly supplied the authors with material for thin sections, and allowed them to quote an unpublished chemical analysis (Table 18). Frankel states that, besides a main acid vein at Poortje, there are also thinner ones which "...diffuse into the dolerite with vague margins and which contain small nests of dolerite, say 1 cm. square". These veins are fresher than the Insizwa examples, and their composition agrees very closely with that of a residual glass separated from a Scottish tholeiite by the senior author (1935, p. 150).

It is unwise, however, to overemphasize chemical resemblances of this nature, for good agreement is also seen with chemical analyses of a suspect microgranite vein from Insizwa (Scholtz, 1936, p. 143) and of mobilized Beaufort sediment from Ibeka store (Mountain, 1944, p. 113). The Kinkell and Poortje analyses are both potassic and doubtless represent the penultimate magmatic residue. The poikilitic biotite of the contact basalt of Downes Mountain and Hangnest was probably formed during the circulation of potassic residual liquids belonging to this stage.

TABLE 18.—*Chemical compositions of residual glass and injection veins*

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O
Kinkell glass	73.0	13.2	1.1	1.6	0.5	2.8	2.6	4.6
Poortje vein	73.2	13.1	0.8	1.6	0.1	2.0	2.5	6.0
Insizwa microgranite	73.9	12.6	1.2	1.2	0.7	1.2	— 8.7 —	
Ibeka mobilized sediment	72.2	14.2	0.1	2.9	0.8	1.9	2.0	3.8

The acid dikes and pipe fillings of the Carnarvon district which have already been described represent mobilization of quartzo-feldspathic material on a much larger scale, but their petrographic resemblance to many of the metasomatic and syntectic rocks of the Karroo renders it improbable that they have escaped contamination. Their relationship to the associated sheets of normal dolerite is not known.

Another locality where acid material has been produced on a large scale is the New Amalfi sheet (Poldervaart, 1944), but the iron-rich dolerites and later acid residues of this highly fractionated intrusion reacted eagerly with the associated Burghersdorp sediments and are therefore greatly contaminated. The same is probably true of the roof zone of the Insizwa-Ingeli intrusion.

The general poverty of acid residual material, contaminated or otherwise, is very striking in the Karroo province. Apparently the later partial magmas were seldom segregated or injected into the solid phase to any great extent, and therefore took the form of interstitial micropegmatite or mesostasis in the normal dolerites. The absence of the late soda-rich, potash-poor veins common in many tholeiitic provinces (Walker, 1940, p. 1093) is also noteworthy, but the feldspathic and microgranitic patches described by Mountain (1943, p. 59) seem to show an approach to this stage of differentiation. Possibly the final residues crystallized in the slightly earlier potash-rich stage, but this seems unlikely, since certain intrusions—*e.g.*, that of the Glen, show marked albitization, which probably took place during a still later hydrothermal stage.

PRODUCTION OF THOLEIITES FROM OLIVINE DOLERITES BY THE INCORPORATION OF RESURGENT VOLATILES

The results of the incorporation of resurgent volatiles from sediments by the dolerite magma were described by the junior author in the Mount Arthur complex (1946). The intrusion is of the bell-jar type, and, due to the collapse of the roof, large blocks of Stormberg sediments were enveloped by the magma (Fig. 8). The inclusions dip in all directions and have sunk in the magma, in some instances over a vertical distance of more than 1000 feet, even in the case of Molteno beds, which

consist of feldspathic sandstones (average sp. gr. 2.66). The normal Kokstad type dolerite of the intrusion gradually gives way to highly altered tholeiites forming mantles round xenolithic blocks. The junior author (1946, p. 104–107) explained these features by a hypothesis involving the incorporation by the magma of resurgent volatiles, distilled from the sedimentary inclusions, and their concentration in a zone immediately surrounding the xenoliths. Local concentration of the volatiles

TABLE 19.—*Average chemical composition of Kokstad type dolerite and Mount Arthur dolerite*

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	H_2O+	H_2O-	P_2O_5
1	50.50	.85	15.55	.87	9.30	.20	8.48	10.75	1.96	.59	.69	.14	.12
2	52.26	1.30	15.14	.96	9.20	.24	6.44	9.70	2.40	.98	.96	.24	.18
3	52.34	.99	14.47	2.33	8.77	.26	6.33	10.25	2.25	.70	1.02	.27	.02

1. Average Kokstad type dolerite (13 analyses).
2. Average of 11 analyses of Karroo dolerites with MgO 6–7 per cent.
3. Average Mount Arthur tholeiite (4 analyses).

resulted in a change in the crystallization trend of the magma, producing tholeiites instead of the normal Kokstad type of olivine dolerite, while it also caused a decrease in the density of the volatile-rich magma. The distillation of resurgent volatiles from the xenoliths is thought to have been a gradual and prolonged process, which may have aided the settling of the included blocks. The petrographical changes of the conversion of olivine dolerite to tholeiite have been noted (Table 3; Fig. 14). Briefly these changes may be summarized as follows:

Olivine first alters to pleochroic serpentine and finally disappears. Pigeonite forms a prominent part of the pyroxenes, while orthopyroxene is absent. The interstitial micropegmatite is replaced by a dark mesostasis, which increases as the included sediments are approached. Hydrous minerals, such as hornblende, serpentine, and chlorite become more prominent and finally occupy vesicles, together with quartz and calcite. Some tholeiites also contain irregular areas of quartz, which show crenulated margins. Alteration becomes more and more intense, affecting plagioclase, pyroxene, and mesostasis alike.

Chemically the tholeiites do not show abnormal relations, apart from a greater degree of oxidation of iron (Table 19).

No great changes in the chemical composition are to be expected, as the addition of hyperfusible components—mainly water—to the magma would result mainly in a change in the course of crystallization of the magma. The higher degree of oxidation of iron, due to the solution of water by the magma, accords with observations made by Phemister (1934, p. 40–41). Hence late iron ore crystallized in greater quantities than in the normal dolerites, while the pyroxene is generally richer in lime (2V:59°–56°) than that of the average Karroo dolerite. The volatiles also promoted the formation of pigeonite, instead of orthopyroxene, as the stable phase at all temperatures. Finally the hyperfusibles were responsible for the high degree of alteration of these rocks and their vesiculation near the xenoliths.

The Mount Arthur complex is not the only intrusion in which the change from dolerite to tholeiite was noted. Along the margins of the New Amalfi sheet (Polder-

vaart, 1944, p. 93) the normal Kokstad type dolerites are also replaced by tholeiites identical with those of Mount Arthur. At Hangnest (Walker and Poldervaart, 1941a, p. 435) tholeiitic modifications of the normal Hangnest or Downes Mountain type dolerite of the sill were found in a wedge above a stoped strip of sediment near the upper contact. Marginal development of tholeiites was noted in other Karroo dolerite sills of Kokstad, Perdekloof, or Blaauwkrans type.

TABLE 20.—*Average chemical composition of Hangnest type and related dolerites*

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
1	52.5	1.0	15.4	1.2	9.3	.2	7.1	10.3	2.1	.8	.1
2	54.4	1.1	14.8	1.3	8.9	.2	7.2	9.0	2.1	.9	.1
3	54.2	1.9	14.7	1.1	8.7	.3	5.8	10.6	2.2	.5	—
4	53.3	.6	16.4	.5	8.3	.2	6.7	11.5	1.6	.9	—
5	51.9	2.4	14.1	4.0	8.7	.2	5.6	9.3	2.5	1.0	.3
6	71.4	.5	14.8	.9	3.2	—	1.4	1.5	3.0	3.2	.1
7	54.2	1.0	15.4	1.2	8.7	.2	6.5	9.5	2.2	1.0	.1

1. Average Karroo dolerite (43 analyses).
2. Average Hangnest dolerite (6 analyses).
3. Bronzite dolerite, Falkland Islands.
4. Average Tasmanian dolerite (6 analyses, Edwards, 1942).
5. Average Whin Sill dolerite (6 analyses, Holmes and Harwood, 1928).
6. Average Beaufort siltstone (6 analyses).
7. Result of adding 10 per cent Beaufort siltstone to average Karroo dolerite.

Hitherto the marginal development of tholeiitic modifications in olivine dolerite intrusions has been attributed to more rapid cooling. This is probably true for those tholeiites which show a decrease in grain size along with the development of mesostasis. Yet in many cases the tholeiites are of equal or even greater grain size than the central olivine dolerites. The hypothesis of the resurgence of hyperfusible components from country rock or xenoliths, and their solution by the magma, throws a new light on the marginal development of tholeiitic phases. At the same time it might explain some observed anomalies in specific gravity relations.

ORIGIN OF THE HANGNEST, HANOVER AND KENTANI TYPES

Hangnest type.—Bronzite dolerites of the Hangnest type are of widespread occurrence in the Karroo, having been recorded in Natal (the Matshongololo type of Blignaut and Furter, 1940), the Transkei, and the Western Karroo. Dolerites of this type are usually younger than the Karroo intrusions of Perdekloof or Blaauwkrans type (Blignaut and Furter, 1940, p. 72; Walker and Poldervaart, 1942b, p. 64). Petrographically they differ from the normal Karroo dolerites by the absence of olivine, while even the chilled modifications carry phenocrysts of bronzite instead of olivine. They show slightly higher SiO_2 and lower Al_2O_3, FeO, and CaO than the average Karroo dolerite (Table 20).

The authors could not satisfactorily determine the origin of this type. It is difficult to explain their chemical composition by normal crystal fractionation and gravitational differentiation of the Karroo magma. Read (1935) has shown that

norites may be produced from olivine gabbros by assimilation of argillaceous sediments. However, there was subtraction, rather than addition, of alumina in the production of the Hangnest type. The addition of 10 per cent Beaufort siltstone to the average Karroo dolerite (Table 20) results in a calculated composition similar to that of the Hangnest type, but too high in Al_2O_3 and CaO and too low in MgO.

TABLE 21.—*Average chemical composition of Hanover type and related dolerites*

	SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
1	52.2	.8	14.9	1.2	8.2	.2	9.8	9.6	2.2	.8	.1
2	52.0	.9	15.9	.8	8.8	.1	8.0	10.6	2.0	.8	.1
3	51.0	1.0	15.5	1.5	9.6	.2	8.4	10.4	1.7	.6	.1

1. Average Hanover type (3 analyses).
2. Average Hanover type, excluding the "chill phase" of Waterfall Gorge (Table 15, no. 35).
3. Average of 3 olivine-bearing chilled basalts (Table 15, nos. 57, 58, 59).

Possibly a combination of fractionation of magma and assimilation of siltstone could account for the production of the Hangnest type, but the exact mode of formation of the type remains obscure.

Comparisons with bronzite dolerites from elsewhere show appreciable differences in chemical composition, though mineralogically the rocks are nearly identical, varying only in the relative proportions of the constituent minerals. The Downes Mountain type of dolerite is invariably associated with the Hangnest type and occupies the higher parts of sills of the latter type. It is a more acid type of dolerite than the associated Hangnest type, owing to an upward migration of the salic partial magmas, combined with slight settling of pyroxene. Together with the salic partial magmas, volatiles, mainly water, became concentrated in the upper portions of the magma. This is shown by the marked elongation and curvature of the pyroxene (Pl. 7, fig. 5), the occurrence of magnesian pigeonite instead of the bronzite of the Hangnest type, and a greater development of biotite and amphiboles. Chemically the change from Hangnest to Downes Mountain type is shown by higher proportions of SiO_2, TiO_2, Al_2O_3, Na_2O, and K_2O, and decreases in the proportions of total iron, MgO, and CaO.

Hanover type.—The Hanover type of dolerite is characterized by the occurrence of large crystals of bronzite, scattered through a matrix of slender plagioclase laths, ophitic augite, small rounded crystals of fayalitic olivine, iron ore, and micropegmatite. Chemically the type has no distinguishing features and closely resembles other Karroo dolerites of the same degree of differentiation, particularly if the somewhat abnormal "chill phase" of the Insizwa sheet is excluded.

It seems clear that no assimilation of sediment was involved in the production of the Hanover type of dolerite. Its origin must be ascribed to physical conditions rather than chemical environment. The composition is that of a dolerite which could reasonably be expected to carry magnesian olivine (Table 21), but the authors believe this olivine was entirely resorbed during the leisurely ascent of the magma, which resulted in abnormally slow cooling. Large crystals of pigeonite were thus

formed at the expense of olivine. After emplacement the magma cooled more rapidly. Pigeonite inverted to bronzite, but lime was not completely exsolved from the crystals, resulting in the curious patchy extinction shown by the bronzite crystals, and demonstrated by the high lime of the analysis (Table 7). The more rapid cooling

TABLE 22.—*Chemical composition of Kentani gap dike rocks*

		SiO_2	TiO_2	Al_2O_3	Fe_2O_3	FeO	MnO	MgO	CaO	Na_2O	K_2O	P_2O_5
South Dike	1	51.5	3.4	15.7	1.1	11.1	.1	3.3	9.1	2.6	1.9	.2
	2	52.6	2.1	16.4	1.0	11.1	.4	3.5	7.9	2.8	1.7	.5
	3	52.9	1.6	16.0	1.5	8.6	.1	5.8	8.9	2.8	1.7	.1
	4	51.8	1.5	15.2	.6	9.8	.2	6.2	11.8	2.2	.6	.1
North Dike	5	64.9	1.7	13.0	.8	7.3	.1	1.2	4.0	3.1	3.7	.2
	6	73.7	.7	14.3	—	2.9	—	1.0	2.1	2.5	2.7	.1
Average Kentani		52.3	2.4	16.0	1.2	10.3	.2	4.2	8.6	2.7	1.8	.3
A.		52.2	2.4	16.5	3.1	9.7	.1	3.5	8.5	3.0	.9	.1

1. Kentani type dolerite (Table 15, no. 36).
2. Kentani type dolerite (Table 15, no. 37).
3. Kentani type dolerite with blotchy appearance (Table 15, no. 38).
4. Dolerite of western end of dike (Table 15, no. 22).
5. Granophyric dolerite, western end of dike (Table 15, no. 85).
6. Feldspathic quartzite, core of dike at western end (Table 15, no. 88).
A. Quartz diabase, 700 feet above lower contact, Palisade sill (quoted from Walker, 1940).

also accounts for the fine grain of dolerites of this type, which is in marked contrast to the large bronzite "phenocrysts". During the later stages of crystallization the liquid returned to the pyroxene-olivine boundary plane, and fayalitic olivine (Fa_{60-90}) crystallized together with iron-rich pyroxene (Bowen and Schairer, 1935, p. 185).

Kentani type.—Opposing opinions have become apparent in the writings of the senior author (1943) and Mountain (1943), who both discussed the origin of the Kentani type of dolerite, occurring in the two gap dikes of the Transkei. As already noted nearly identical rocks are found as dikes in the Northern Karroo (Rogers, 1906, p. 66–67); but they have pink instead of white orthoclase.

Mountain studied the gap dikes along their entire lengths and made many new observations, in particular as to their field relations. These are briefly summarized here:

Southern dike, 96 miles long, 300 yards wide, narrowing rapidly toward the west and finally disappearing near Cathcart. When traced toward the west the rock becomes more and more blotchy in appearance and eventually passes into a thin dike of Karroo dolerite. In places the normal dike rock becomes more feldspathic and paler. Feldspathic and micropegmatitic segregations in the form of schlieren are found in an exposure at Blue Lagoon, just north of the Qora River mouth. A xenolithic dark patch, resembling ordinary dolerite, but rich in biotite, is also found here.

Northern dike, 93 miles long, 210 yards wide, tapering out toward Cathcart. Toward the west the rock becomes finer-grained and more granophyric. In the neighborhood of Cathcart there is a core, about 7 yards wide, of quartzitic material, which closely resembles the indurated sandstone of the country rock, flanked by margins of pale granophyric dolerite.

Both dikes are younger than the Karroo dolerite sills, occurring in the district. Mountain (1943, p. 57) also notes that

"...while the greater part of the dikes in cross-section may be fairly homogeneous, it is not uncommon to find a selvage, sometimes several yards wide, of finer grained and more basic material. Thus in the Transkei, the margin is often a more or less normal dolerite as seen in the field, while the bulk of the dike is a coarse leucogabbroid rock".

Relevant chemical analyses are reproduced in Table 22.

TABLE 23.—*Micrometric analyses of gap dike rocks*

	1	2	3	4
Plagioclase	54	63	—	—
Albite	—	—	66	38
Orthoclase	—	6	20	22
Pyroxene	23	7	—	—
Biot., Amph., Serp.	15	12	—	5
Iron Ore	4	2	—	—
Micropeg. and Qu.	4	10	14	35
Color index	42	21	0	5
Sp. gr.	2.93	2.72	2.61	2.58

1. Normal gap rock.
2. Pale modification of gap rock.
3. Feldspathic schlieren.
4. Micropegmatite schlieren.

No chemical analyses of the interesting schlieren in the southern dike at Blue Lagoon are given, but Mountain determined their mineral composition with a Dollar stage. These volumetric data are reproduced here in Table 23.

Mountain (1943, p. 67) considers it possible that "...the dolerite pegmatite magma forming the greater part of the gap-rock was generated by association in depth of the normal Karroo dolerite magma with deeply buried sediments". The doleritic margins of the dikes are thought to represent "...a chill phase of the normal gap-rock magma". He also thinks it possible that the segregation schlieren in the southern dike "...are really the remains of incorporated sediments that have derived their soda from some hydrothermal residue of the dolerite magma". The progressive acidification of the dike from east to west is considered "...due to increasing contamination of the dolerite pegmatite magma by incorporation of sedimentary material from the invaded Beaufort sandstones and mudstones".

The present authors believe that the gap dikes more probably represent normal products of fractionation of the Karroo magma (*cf.* Walker, 1943); their reasons are as follows:

(1) In chemical composition the average Kentani type is very similar to a quartz diabase from the upper part of the Palisade sill. The average Palisade diabase shows a very close chemical correspondence with the average Karroo dolerite (Table 17), and the Palisade magma was notably inert in its behavior toward the associated sediments.

(2) The composition of the average Kentani type reflects all the changes expected

in the fractionation of the Karroo dolerite magma. Thus the relative impoverishment in MgO and CaO and the enrichment in FeO, Fe_2O_3, TiO_2, Na_2O, and K_2O, when compared with the average Karroo dolerite, are explained by normal fractionation of the Karroo magma.

(3) Mineralogically the Kentani type resembles the dolerite pegmatites of the Karroo, though the latter are generally more mafic. There is no need for an appeal to metasomatism of sediments to explain the origin of dolerite pegmatites, which were considered by Tomkeieff (1929) to be due to the segregation of aqueous portions of the magma.

(4) Metasomatic granophyres of the Karroo are characterized by an elongated habit of the pyroxene generally pseudomorphed in yellow-green serpentine (Pl. 10, fig. 1), and by the occurrence of relict minerals and structures. The Kentani type dolerite bears no resemblance to these granophyric rocks of demonstrably metasomatic origin.

The authors consider then, that the Kentani type most probably represents an advanced stage of differentiation of the Karroo magma, emplaced after the main period of injection of Karroo dolerite. The "acid" plutons of the highly differentiated Lebombo region (Du Toit, 1929b) form a comparable example in South Africa. Mountain's field observations are of value in that they allow a more complete picture to be drawn of the emplacement of the gap dikes.

The "acid" magma had apparently reached its maximum iron enrichment when emplaced, and differentiated rapidly toward the alkaline and siliceous residue, which characterizes the final stage of crystallization of basalt magma. Thus segregations of the dike consist of feldspar or micropegmatite and are poor in ferromagnesians. The mineral composition of the segregations of Blue Lagoon agrees very well with expectations, hence a sedimentary origin for these patches is considered improbable. (*Cf.* Mountain, 1943, p. 67.) Apparently the "acid" Kentani magma was exceedingly mobile and active in its behavior toward the associated rocks. It followed the course of two earlier dolerite dikes, remnants of which are still preserved along the margins of the dikes. The original dolerite was converted to granophyric dolerite, rich in biotite, by normal reaction with the Kentani magma, in accordance with the reaction principle (Bowen, 1928, p. 197–201).

The northern dike becomes more acidic when traced toward Cathcart, this being ascribable to further differentiation of the Kentani magma or to contamination by Beaufort siltstone. At Stoneridge a core of feldspathic quartzite occurs in the dike, flanked by granophyric, contaminated dolerite. There can be little doubt, in view of the evidence cited by Mountain (1943, p. 70–71), that this core represents syntectic material, derived from Beaufort siltstone, and affected by emanations from the earlier dolerite, from the later Kentani magma, or from both. The occurrence of the syntectic core at Stoneridge is not necessarily evidence for a metasomatic origin of the two gap dikes, as implied by Mountain.

BASALTIC MODIFICATIONS OF KARROO DOLERITES

General statement.—Two types of basaltic rock are associated with the Karroo dolerites. The first is very common and consists of the chilled contact rock which

lines the vast majority of the intrusions. The second is much rarer and takes the form of veins and dikes intruded into earlier dolerite intrusions of normal character.

Chilled contact basalts.—The vast majority of Karroo dolerite intrusions are fringed by a thin layer of basaltic rock which is generally, but not invariably, continuous, and it would seem that the country rock at the time of emplacement was everywhere at a sufficiently low temperature to cause some chilling. The contact basalt grades into dolerite, and in most cases only a few inches are of a grain size fine enough to warrant the use of the former term. But in spite of its thinness the basaltic rock doubtless acted as an ill-conducting couche which both preserved the heat of the magma and also insulated it from contact with the country rock, with which it would have otherwise reacted. In a few cases the basaltic contact has been stripped off, principally from the upper contact of horizontal or gently inclined intrusions in which differentiation has been relatively great with the production of a highly reactive roof zone—*e.g.*, at New Amalfi (Poldervaart, 1944, p. 87). The upper chilled phase may then be distributed through the higher portions of the sill as xenolithic blocks. In other cases marked chilling at the upper contact gives place to transfusion processes which have produced a gradation from dolerite to country rock, as at the inclined sheet of the Glen (Walker and Poldervaart, 1942b, p. 63).

Besides preserving the magma from reaction with the country rock, the basaltic chilled phase gives a check on the chemical and mineralogical constitution of the magma at the beginning of emplacement. It must be recognized, however, that the composition of the magma may change during emplacement, or that, after it has become static, differentiation or reaction with sediment may take place so that the basaltic contact phase may show marked differences in composition from the main portion of the intrusion. An inclined sheet at Thee Kloof has a basaltic contact containing phenocrysts of olivine and basic plagioclase, and a center consisting of olivine-free bronzite dolerite of Hangnest type but this may be a composite intrusion (Walker and Poldervaart, 1942b, p. 64). Wahl (1946) has pointed out that composite dikes and sheets may originate by a combination of thermal diffusion and convection. In other cases—*e.g.* at Hangnest (Walker and Poldervaart, 1941a, p. 432)—there is evidence of slight changes in the chemical and mineralogical constitution of the chilled phase by the circulation of late potassic solutions which have precipitated poikilitic biotite as a reaction product. In general, however, the contact basalts are little altered and represent accurately the constitution of the magma when first intruded.

An interesting comparison may be made between the average chemical compositions of the contact basalts, the undifferentiated dolerites, and the Stormberg basalts (Table 24).

The three averages show remarkable agreement apart from the relatively oxidized state of the iron in the Stormberg lavas and slightly lower magnesia and higher alkalies in the dolerites and Stormberg lavas. These small differences may be due to some settling of olivine, though the constancy of alumina does not support this. The conclusion may be drawn that a very considerable number of dolerites and probably an equivalent proportion of basaltic lavas in the province have undergone

little change in composition since emplacement or extrusion took place. Thus the average chemical compositions of the chief types of dolerite are grouped quite closely around the grand average (Table 16).

The proportion and make-up of the phenocrysts in the contact basalt is of considerable importance since they indicate the amount of the solid phase present in the ascending magma and also the minerals which had commenced to crystallize at that stage.

TABLE 24.—*Average chemical composition of lavas and contact basalts*

	SiO_2	Al_2O_3	Fe_2O_3	FeO	MgO	CaO	Na_2O	K_2O
1	51.9	15.5	1.1	9.7	8.2	9.7	1.8	0.7
2	52.5	15.4	1.2	9.3	6.9	10.4	2.1	0.8
3	52.2	15.6	2.7	7.9	7.2	10.1	2.0	0.9

1. Average contact basalt (6 analyses).
2. Average undifferentiated dolerite (43 analyses).
3. Average Stormberg lava (5 analyses).

In general, all the earliest of the main constituents of the associated dolerite are represented, and cases where basalt and dolerite show appreciable mineralogical differences are rare. Olivine dolerites thus show phenocrysts of magnesian olivine, augite, and basic plagioclase in the contact phase; the texture becomes progressively seriate as the grain size increases. Similarly, bronzite dolerites of Hangnest type have phenocrysts of bronzite and basic plagioclase and occasionally of subordinate augite.

The common occurrence of augite among the phenocrysts of the chilled phase is not easily explained but is a problem encountered in other tholeiitic provinces—*e.g.*, the Whin Sill and Central Scotland. Augite may crystallize independently or associated with plagioclase as portions of glomeroporphyritic aggregates. Most of the Karroo dolerites are definitely ophitic, and augite would seem to be of late crystallization, unless it is the only pyroxene present as in fine-grained dolerites of Perdekloof type. This mineral has, however, a long range in the crystallization sequence and is one with low powers of spontaneous crystallization, so that it may have commenced to crystallize relatively early. Moreover, under intratelluric conditions it may have started to crystallize still earlier, since increased pressure produces a sharper rise in the melting point than in the case of the other essential constituents.

Some at least of the bronzite phenocrysts in the chilled margins of dolerites of the Hangnest type may be reaction products derived from still earlier magnesian olivine, for they occasionally enclose small rounded relics of the latter mineral.

Table 4 shows that the proportion of phenocrysts in the marginal basalts attains a maximum of 15 per cent in the type Perdekloof sill but is generally much lower; the average is only 6.7. Crystallization cannot, therefore, have been far advanced when the Karroo dolerite magma was emplaced. The average proportion of true

phenocrysts may be lower than the figures quoted in Table 4, for some of the glomeroporphyritic aggregates may be xenocrystic. Of the total average of 6.7 per cent, plagioclase forms 4.2, and augite only 0.8 per cent.

Late basalt veins and dikes.—Veins and dikes of basaltic or tachylitic rock in normal dolerite have been recorded from several parts of the Karroo, but only those in the neighborhood of Aliwal North have been dealt with thoroughly (Frankel, 1942; 1943). In some cases they appear to be of similar composition to the normal rock, but those described in detail by Frankel are distinctly more basic, being rich in olivine and containing picotite. Some show chilling against the normal dolerite, while others are unchilled.

The phenomenon is fairly common in dolerite provinces and besides the examples referred to by Frankel (1943, p. 53) has been recorded by Geikie (1897, vol. 2, p. 310) in the Shiant Isles of the Hebrides, and by the senior author in Central Scotland (1935, p. 139, 146). The only satisfactory attempt to explain the origin of these veins is that of Frankel (1943, p. 52) who suggests that

> "... the crystallisation of a differentiated coarse-grained basic phase in the magma chamber was followed by the implacement of the greater mass of magma through overlying strata. A disturbance in the magma chamber probably caused by faulting, resulted in the break-up of the basic phase, fragments of which were carried upwards either in viscous magma still in the magma chamber, or in magma formed by the reheating of portion of the main rock body".

This explanation would also account for later injections of the same composition as the normal rock if the original magma chamber were not differentiated. The reheating hypothesis advanced by Frankel would involve very considerable rises of temperature, for the glassy nature of the basaltic veins and dikes indicates that they were not merely mobilized but fused to a very considerable extent. In the case of certain more acid veins of basalt glass near Durban, now being investigated by Frankel and Kent (Table 15, analyses 63 and 64), the melting was still more complete. Clearly the problem requires further investigation.

Besides the basaltic veins cutting normal dolerite, others have been noticed by the senior author at Umhlali Beach (north of Durban) passing from the normal dolerite into the adjacent sediment.

REACTION OF DOLERITE MAGMA WITH SEDIMENTS

GENERAL STATEMENT

The extreme activity of the Karroo dolerite magma toward the associated sediments has been emphasized. Besides the hornfelses and hornstones which are the normal products of pyrometamorphism under hypabyssal conditions there are countless cases where the country rock has been mobilized or metasomatized, and a few where something like assimilation or pure melting has taken place. The processes involved may be classified as follows:

Rare
(1) *Fusion of sediments with the production of glass*
(2) *Assimilation of sediments*

Common

(3) *Transfusion or metasomatism of sediments—i.e.*, the molecular replacement of the original minerals by new minerals through the agency of emanations from the dolerite magma. The replacement is sufficient to cause marked alteration in the appearance and composition of the sediment and the product is frequently igneous looking. In the most extreme case the altered material may become rheomorphic and vein the dolerite. Transfusion of sediment is generally accompanied by a corresponding change in the appearance and composition of the adjacent dolerite.

(4) *Rheomorphism of sediments.* This term covers all processes whereby sediments are fused sufficiently as to become capable of flow. Rheomorphism may be anatectic or syntectic and may cause brecciation of solid dolerite. Veining of dolerite by mobilized country rock is very common.

The abundance of (3) and (4) in the Karroo dolerite province is one of its most interesting yet puzzling features. Such abundance would be understandable if there had been great superheat or unusually high magmatic temperature, but no evidence of either is found. On the contrary the paragenesis of the pyroxenes suggests that the temperature of the magma during emplacement was between 1100°C and 1150°C. The rarity of tridymite among the baked quartzose sediments is noteworthy. Although the inversion from quartz to tridymite (870°C) is sluggish, tridymite is not uncommon in the baked sediments of other dolerite provinces (see Reynolds, 1908, pl. LII, fig. 6; Thomas, 1922, pl. VII, fig. 5; Campbell and Stenhouse, 1934, p. 166). The degree of pyrometamorphism of the country rock in the Karroo does not indicate any great superheat in the magma though it may be considerable in dolerites of Hangnest type (Blignaut and Furter, 1940, p. 67).

We are forced, therefore, to one of three conclusions:

(1) The magma may have been abnormally rich in volatiles;

(2) fluxes (*e.g.*, chlorides and sulphates) may have been unusually abundant in the country rock;

(3) physical conditions were very favorable for the transfusion process.

VOLATILE CONTENT OF KARROO DOLERITE MAGMA

Water is the chief volatile in basaltic magma, but there are wide differences of opinion as to the amount present. Bowen (1928, p. 301) considers that 0.5 per cent represents the order of magnitude, but Phemister (1934, p. 35) takes 5 per cent as a more probable value under deep-seated conditions. The authors follow Daly (1944, p. 1378) and Shand (1943, p. 39) in accepting the higher estimate and consider that the Karroo dolerite magma was normal in its water content. The mineralogy of the Karroo dolerites agrees with that of other tholeiitic provinces, and there is no special abundance of the pegmatitic schlieren, patchy segregations, and residual injection veins which characterize magmas of high water content.

The only direct evidence as to the amount of water present in the magma is given

by two fresh glassy veins investigated by Frankel and Kent (1948). One of these from Verulam represents relatively unfractionated Karroo magma and contains 3.67 per cent of H_2O^+. The second glass from Durban belongs to a later stage in the differentiation sequence and has an H_2O^+ value of 8.73 per cent. Therefore, the water content of unfractionated Karroo magma was probably of the order of 5 per cent, though this value would rise considerably during crystallization and might exceed 10 per cent during the period of maximum transfusion. Thus the available evidence points to a normal volatile content for the Karroo dolerite magma, and other reasons must accordingly be sought to explain its unusually active behavior toward the associated sediments.[8]

HYPERFUSIBLE COMPONENTS OF THE KARROO SEDIMENTS

As there is no evidence of an abnormally high volatile content in the Karroo dolerite magma the hyperfusible content of the associated sediments must next be considered, and a difficult problem is immediately encountered. Rheomorphic and metasomatic phenomena have been brought about by the dolerites not only in all divisions of the Karroo System but also in the underlying Table Mountain Series. The original water content of the strata involved must have shown wide differences for they include all gradations between argillaceous and arenaceous [*e.g.*, Table Mountain sandstone (Mountain, 1935), Dwyka tillite (Du Toit, 1912), Ecca and Beaufort siltstones (Walker and Poldervaart, 1942a), Molteno grit (Poldervaart, 1944), Cave Sandstone (Du Toit, 1904, p. 176–177)]. Possibly fluxes such as SO_4 and Cl may have played a part in the reactions though the variability of the rock types affected would indicate that they were introduced by ground water in some cases rather than occurring as original constituents of the sediment. Such fluxes would undoubtedly promote rheomorphism and transfusion of sedimentary material, and the universally high salinity today of Karroo waters (Bond, 1946) may well be an indication of their presence in the strata concerned during Trias-Lias times.

Tests were made by the authors on Beaufort siltstone (a type peculiarly prone to transfusion phenomena) from freshly blasted exposures above the water table. The rock contained 2.3 per cent H_2O^+ but only traces of SO_4 and Cl. This does not prove that the fluxes were not present formerly in greater abundance, for both are highly soluble, but if the sediments have been leached the process must have been rapid in view of the present low rainfall (8 inches) of the region. Slices of the same sediment when heated to 1100°C in an electric furnace in a CO_2 atmosphere showed only slight flexing of the unsupported portions. The petrological and field evidence indicates that the rheomorphic siltstone veins were injected at temperatures well

[8] The Edinburgh district of Central Scotland provides one of the few demonstrations that the volatile content of basaltic magma has an important bearing on its power to react with sedimentary strata. Two suites of dolerites were injected during Carboniferous time into the same series of Lower Carboniferous sandstones and shales. The earlier suite is teschenitic and is relatively rich in biotite, hornblende, and zeolites as well as the schlieren, patchy segregations, residual injection veins, and other phenomena characteristic of aqueous magmas (Flett, 1929; 1930; 1931). The later suite, on the other hand, is tholeiitic and was clearly of lower water content for it is relatively free from hydrous minerals and from "wet" segregations. The earlier volatile-rich suite shows striking examples of reaction with the associated sediments, while the later tholeiitic one is completely inert toward them. Two fresh undifferentiated, glassy dolerites from the later suite have H_2O+ percentages of 2.50 and 2.70 (Walker, 1935, p. 140) which probably means an original content of the order of 5 per cent since the vesicular margins of both intrusions indicate some escape of volatiles.

below 1000°C after the solidification of most of the dolerite, which is again indicative of changes in the composition of the original sediment—probably through leaching.

The Karroo Basin must have formed a perfect artesian structure before emplacement of the dolerites, allowing brackish waters to accumulate in the central portion, but this does not explain the exceptional abundance of rheomorphic and transfusion phenomena found in the marginal regions of Natal and the Transvaal. All considered, the ground waters of Trias-Lias times were probably rich in fluxes which would promote rheomorphism and transfusion (by means of dolerite intrusions) in the sediments through which they circulated.

PHYSICO-CHEMICAL ASPECTS OF THE PROBLEM

The abundance of reaction phenomena between Karroo dolerite magma and the adjacent or included strata is quite abnormal and cannot be explained adequately by high water content or the presence of fluxes in the sediments. Other normal tholeiitic provinces are associated with similar strata yet show no comparable record of transfusion or rheomorphism. Therefore, physical and chemical conditions were probably at an optimum for the initiation of these processes during the cooling of the intrusions after emplacement.

Data as to the physical conditions favoring transfusion are scarce. Morey (1922, p. 226) has shown that:

> "as a magma containing water and other volatile components cools, with consequent crystallization, the pressure will rapidly rise from its initial value, and as the cooling continues the pressure will increase until the temperature of maximum pressure has been reached, or until the pressure is relieved by the escape of the volatile material".

During the crystallization of basaltic magma with an initial H_2O content of 5 per cent there would accordingly be an increase in both vapor pressure and water content until the latter had reached a maximum. This maximum has been shown by Goranson (1931, p. 481) to be about 9 per cent for granitic magma and is probably of the same order of magnitude in the case of basaltic magmas such as the Karroo. Data concerning the vapor pressure of the magma at this stage of its crystallization are also scanty, but an estimate of 2000–3000 atmospheres seems reasonable (Morey, 1924; Goranson, 1938, p. 272).

According to Shand (1944, p. 348) further crystallization must lead to one of the following consequences:

(1) Excess of water may escape by diffusion into the country rock.
(2) Excess of water may enter into the solid phase as a constituent of hornblende or biotite.
(3) Excess of water may segregate as pegmatitic schlieren in the manner suggested by Tomkeieff (1929).

The first is by far the most important in the Karroo province. Hornblende and biotite are scarce in the dolerites, and pegmatitic schlieren are by no means abundant.

The penetration of aqueous partial magmas into country rock has been discussed by Graton (1940) and by Niggli (1929) in connection with ore formation. Some of their conclusions are applicable to the Karroo. Insufficient physical and chemical

data preclude a detailed picture of the transfusion process, but a few of the main features may be outlined.

The process seems to have begun after most of the dolerite had solidified—probably during the stage of maximum iron enrichment when vapor pressure and chemical energy of the partial magmas were near their peak. It is believed to have continued

TABLE 25.—*Relation between depth and static load in Karroo beds*

Horizon (Base of formation)	Depth below surface (Feet)	Static load (Atmospheres)
Stormberg lavas	4,500	400
Stromberg sediments	8,900	850
Beaufort Series	17,900	1700
Ecca Series	23,900	2250
Dwyka Series	27,800	2550

throughout the period of alkali enrichment—*i.e.*, after about 90 per cent of the original magma had crystallized and the partial magma was undergoing rapid changes of composition. Presumably the relations between static load and vapor pressure of magma favored penetration of country rock by partial magma at this period. Table 25 shows the static load for various horizons in the Karroo System. In the great majority of Karroo dolerite intrusions the static load was enough to prevent vesiculation of the margins.

Not only did the balance between static load and vapor pressure favor penetration of the country rock by the magmatic emanations, but the contact sediments must have allowed easy penetration by the emanations yet not too easy to permit an overrapid escape into regions beyond the zone of reaction.

The magmatic liquids are thought to have mingled with the pore fluid of the country rock (developed from water or fluxes such as Cl and SO_4 which it might contain), the mixture providing a medium of low viscosity through which diffusion could take place readily. The high water content of this medium would promote ionization and break down some of the complex silicates of the partial magma thereby accelerating diffusion still further. Conditions would thus favor formation of new minerals and the transfusion process generally.

If the proportion of the fluid phase in the country rock became sufficiently great, mobilization and rheomorphic injection would result. MacGregor and Wilson (1939, p. 210) estimate that "an ideal mass composed of spherical grains would become mobile if 26 per cent of its volume were liquid". In the Karroo sediments mobilization would be produced by a smaller proportion of the liquid phase for the sediments are polygenetic and generally show good gradation of grain size. Anatectic phenomena are most abundant in argillaceous sediments which contain a high proportion of water and would thus become mobilized relatively easily.

It has already been emphasized that transfusion and rheomorphism of the contact sediments are associated with a late stage in the crystallization of the dolerite magma. At Hangnest and Alewyn's Gat there is evidence that rheomorphic material from stoped blocks of siltstone brecciated not only chilled contact basalt but dolerite

itself (Figs. 10, 38; Pl. 9, fig. 5). The magmatic fluids bringing about transfusion must, therefore, have made their way through contraction joints and other division planes in the already solid basaltic contact phase. Further jointing during the actual period of rheomorphic injection seems to have pinched out and dislocated veins of plastic siltstone in at least one case (Pl. 6, fig. 2). Probably the rheomorphic veins of sedimentary material followed in general the channels already traversed by the magmatic fluids, though evidence of this is difficult to obtain.

Conditions particularly favor transfusion and rheomorphism in the case of xenolithic blocks and of sediments either bordering the upper contact of sills or acting as partings, for here the strata are either surrounded by magma or immediately above the zone in which the volatiles tend to concentrate.

CHEMICAL DATA

Discussion.—The chemical reactions involved in transfusion are obscure and can only be outlined at present owing to lack of analytical data. The first difficulty—a common one in transfusion problems—is variation in the composition of the sediments. Although the Karroo sediments are in no way abnormal in this respect, a single analysis cannot be considered representative of the bulk average composition of the strata. Also the analyzed sediment may already have undergone some metasomatism.

The other end point—normal Karroo dolerite—is much more constant in composition. But the transfusion process is brought about not by the undifferentiated magma but by a highly fractionated residue undergoing rapid changes of composition —a stage of iron enrichment being followed by enrichment in alkalies. Potash was then eliminated from the magmatic residue leaving a final fraction rich in soda. It is thought that magmatic emanations causing transfusion corresponded in chemical composition to the particular stage of differentiation at which the process took place, but transfusion by sodic emanations appears to have preceded the potassic phase owing, doubtless, to the more rapid diffusion of the former.

The composition of the magmatic emanations can only be guessed, but analyses of differentiated dolerites and of residual injection veins may give valuable indications. The chemical composition of the pore fluid of the sediments is even more problematical though a high water content and the presence of fluxes such as Cl and SO_4 is probable.

The combined sedimentary pore fluid and magmatic emanations sought to establish phasal equilibrium with the solid material through which they circulated, and reciprocal reactions followed. Their chemistry is by no means clear, but Mg and Fe seem to have been added to the sediment while Si was added to the transfusing liquid. Other constituents show irregular variations in accordance with the particular stage of fractionation at which transfusion took place.

In Table 26 additions and losses of the chief oxides during transfusion are calculated. The figures are only approximate as the sediments are distinctly porous, and hence constant volume calculations are unreliable. To compensate for the porosity of the rocks, calculated norm specific gravities (Iddings, 1920) were used instead of the determined specific gravities shown in Table 15.

The following are good examples of transfusion which have been studied in detail.

Iron-rich phase of emanations.—Though Table 26 shows that Fe was added in every instance this is most marked at New Amalfi. Here a thick differentiated sheet of Kokstad type dolerite has transfused a sandstone xenolith stoped from the

TABLE 26.—*Addition and loss of chief oxides during transfusion*

Index (Numbers of Table 15)	New Amalfi 108–92		New Amalfi 108–93		Thomas River 106–85		Alewyn's Gat 102–90		Hangnest 104–94		Rietkop 101–96		Rietkop 101–95		Gonubie Source 107–99	
	+	−	+	−	+	−	+	−	+	−	+	−	+	−	+	−
SiO_2		34.0		32.4		16.9		11.7		9.6		1.7		10.9		11.5
TiO_2	3.7		3.7		3.4		.5		.2			.2	—		.4	
Al_2O_3		3.9	5.5			3.4		1.9	.6		.7		2.4		8.6	
Fe_2O_3	6.4			.6	1.2		5.1		1.0		.2		2.2		—	
FeO	21.7		14.2		13.5		.1		3.2		1.4		3.2		3.4	
MnO	.6		.4		.3		.2			.2	—		—		.1	
MgO	1.0		4.5		1.4		4.9		.8		—		1.9		3.5	
CaO	13.0		10.0		8.6		11.9		3.3			.8		1.9	.6	
Na_2O	2.1		8.1		.9			1.0	2.5		2.1			1.2		3.7
K_2O		3.2		5.1		.2		4.3		—	2.4		4.9		3.2	
P_2O_5	.9		.3		.5		.1		.1			.1		.1	—	

upper contact (Poldervaart, 1944; Fig. 7). Kernels of unaltered sandstone are surrounded by transition zones, and all contacts are sharply defined. Abnormally coarse granophyre adjacent to vertical joints may show rude grid-iron structures due to interlocking ferromagnesian minerals. Analyses show marked desilication of the sandstone accompanied by additions of Fe, Ti, and Ca. The Thomas River analyses are comparable to those of New Amalfi (Table 26).

At Alewyn's Gat, in a road cut 10 miles north of Beaufort West, a lenticular siltstone xenolith (50 feet by 8 feet) is enclosed in a 600-foot sill of Perdekloof type dolerite about 100 feet from the sharply transgressive upper contact (Walker and Poldervaart, 1942a; Fig. 37). The xenolith is surrounded by a 2-foot shell of granophyre (Pl. 10, fig. 1) and all contacts are sharp. Calcareous nodules in the sediment have been converted to brownish-pink garnet in the granophyre. Above and especially below the xenolith occur schlieren of dolerite pegmatite, while rheomorphic veins emerge from the granophyric shell, one of them showing subangular patches of brecciated dolerite. Analyses show additions of Fe, Mg, and especially Ca, with losses in Si and K. Iron enrichment at New Amalfi was more pronounced than at Alewyn's Gat and addition of Ca in the latter case overshadows that of Fe.

Hence it seems that transfusion effected by early emanations from crystallizing dolerite magma results in the addition of Ti, Fe, Mg, and Ca to the affected sediments, accompanied by their loss of Si, Al, and K to the metasomatizing fluids. The effect is one of basification of the sediment.

Sodium-rich phase of emanations.—The most marked addition of Na occurred at New Amalfi (Table 26), indicating that soda-rich emanations issued from the partial dolerite magma shortly after the iron-rich emanations. It was accompanied by

addition of Al, in contrast to the initial removal of that element by the iron-rich emanations.

Hangnest (Calvinia) forms another example of soda enrichment. Here a detached siltstone block stoped from the upper contact of a 500-foot sill is surrounded by Downes Mountain type dolerite and has been converted to granophyre (Walker

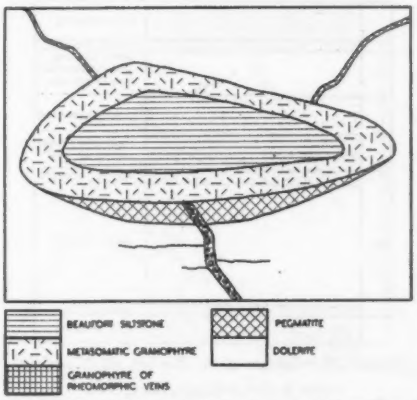


FIGURE 37.—*Sketch of the xenolith at Alewyn's Gat (Beaufort West)* (After Walker and Poldervaart, 1942a).

and Poldervaart, 1941b; Fig. 10). A siltstone block on the same strike line is only partly detached and has not been transfused, but shows rheomorphic and brecciation phenomena (Pl. 6, fig. 1; Pl. 9, figs. 4, 5). Analyses show additions of Fe, Ca, and Na.

Hence the second phase in the transfusion process appears to be one of albitization, resulting chiefly in the addition of Na and Al, removed previously by the iron-rich emanations. If superimposed on the first phase it seems likely that some Si was also added, after its initial subtraction.

Potassium-rich phase of emanations.—The final phase of potassium enrichment is seen at Rietkop, where a 200-foot sill of Hangnest type dolerite has transfused a 70-foot parting of Ecca siltstone to granophyres of variable grain-size, an upper coarse-grained portion overlying one of finer grain (Walker and Poldervaart, 1942a; Fig. 38; Pl. 5, fig. 2). The fine-grained granophyre contains a persistent band in which rounded kernels of unaltered siltstone are conspicuous (Pl. 10, figs. 2, 3). The dolerite below the parting is markedly decomposed and acidified, and all contacts, including that between the two granophyres, are sharp. Rheomorphic veins issue from the upper side of the parting. The Rietkop analyses show that the addition of K was accompanied by loss of Na added during the soda-rich phase of emanations. Addition of K and loss of Na is also shown by the Gonubie Source analyses. There can be little doubt that Si was added during the final phase of the transfusion process, after its initial removal during earlier phases. The effect is, therefore, one of granitization.

Other examples.—The above examples cover all three phases of transfusion, but other interesting cases of metasomatism and mobilization have been described in detail by Mountain from Coedmore Quarries near Durban (1935) and Gonubie River Mouth (1936; 1944). According to Mountain (1943) the gap dikes of the Transkei represent metasomatism and mobilization of sediments on a much greater

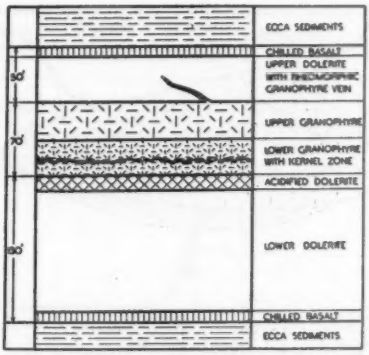


FIGURE 38.—*Section of the Rietkop sill* (*Calvinia*)
(After Walker and Poldervaart, 1942a)

scale than in any of the above examples, but they may be due, in part at least, to crystallization differentiation at depth (Walker, 1943).

The 600-foot inclined sheet of the Glen north of Graaff Reinet shows a gradational contact with the adjacent hornstone which carries abundant titanaugite at the contact (Walker and Poldervaart, 1942b).

Small scale phenomena of metasomatism and rheomorphism at an unusually accessible locality are exposed along the upper contact of an 80-foot inclined sheet on the east side of Beaufort West Dam (Pl. 6, figs. 2, 3; Pl. 10, fig. 4).

General observations.—The implications of the widespread transfusion of sediments by Karroo dolerite are of some importance. Although magmatists believe that large bodies of granitic rocks may be produced by crystallization differentiation of basalt magma, many cogent objections have been raised against this hypothesis (*e.g.* Holmes, 1936). In the Karroo comparatively small intrusions of basalt magma without superheat have produced significant bodies of igneous-looking rocks of granitic composition by transfusion and sometimes mobilization of the adjacent sediments. It is thus possible that transfusion was operative elsewhere, and produced the associations of basic and acid rocks which are so wide-spread.

The sharp internal contacts, the absence of basaltic chilled phases, and the rheomorphic veins and dikes of granophyre in dolerite create a false impression that the Karroo granophyric bodies represent later intrusions of acid magma. Moreover, liquid immiscibility cannot explain the association of basic and acid rocks in the Karroo, since the one group was solid when the other was irruptive.

A sharp distinction between solid diffusion, metasomatism, transfusion, and rheomorphism proved impracticable for the Karroo. Possibly initial changes were effected by solid diffusion, but it seems clear that the major changes were produced

by a fluid medium, comprising both magmatic emanations and sedimentary pore fluid. There was probably a gradual increase in the liquid phase of the transfused sediment as the energy level rose, the partial dolerite magma transgressing into the sediment, and the more refractory elements of the sediment dissolving in the pore

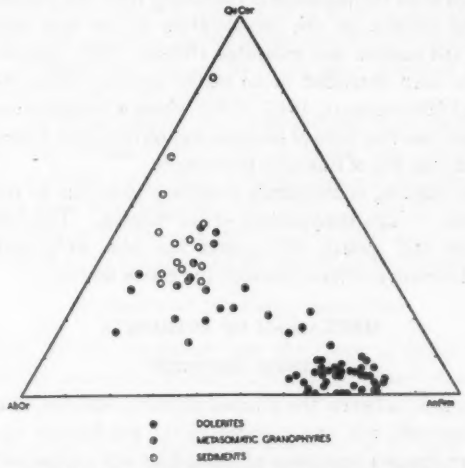


FIGURE 39.—*Analyses of Karroo dolerites, dolerite pegmatites, granophyres and sediments* Plotted on a triangular diagram with variables An Fem; Qu Cor; Or Ab.

fluid. Rheomorphism of the granophyre is thus regarded as the culmination of the transfusion process. Before transfusion the liquid phase was centered in the dolerite intrusion. During rheomorphic injection of dolerite by granophyre the liquid phase was centered in the transfused sediment. During the intermediate stages—*i.e.* during transfusion—there was no distinction between residual dolerite magma and sedimentary pore fluid, other than controlled by environment and diffusion. The transfusion process was marked by a falling energy level in the dolerite with rising energy level in the sediments.

The aqueous medium, and the substances dissolved in it, reacted both with the solid phase of the dolerite and of the transfused sediment. The liquid followed the normal differentiation trend of the magma, an iron-rich phase being followed by a soda-rich phase, and finally by a potash-rich phase. Initially the liquid gained felsic constituents (Si, Na, K, Al) from the sediment giving in return mafic constituents (Ti, Fe, Mg, Ca). At a later stage the now highly acid fluid attacked both the basified sediment and the adjacent dolerite. The changes effected in the transfused sediment can be divided into (1) basification, (2) albitization, and (3) granitization, comparable to the sequence of changes recognized by Reynolds (1946) in the granitization process.

Analyses of Karroo sediments, dolerites, dolerite pegmatites, and metasomatic granophyres are plotted on a triangular diagram with variables An Fem; Qu Cor; Or Ab (Fig. 39). The figure illustrates the tendency of the transfused sediments and the dolerite differentiates to approach compositional uniformity.

Assimilation and pure melting.—The widespread transfusion of sediments by Karroo dolerite is in marked contrast to the rarity of assimilation phenomena. Those which do occur belong to an early stage in the cooling history of the magma which seems to have had little superheat. The felsic Karroo sediments were later in the reaction series than the products crystallizing from the magma and could only be dissolved by the magma by the precipitation of the heat equivalent of those phases with which the magma was saturated (Bowen, 1928, chapter X). Examples of the process have been described in an earlier section, while other instances are seen at Mount Fred (Poldervaart, 1944, p. 90) where a hornstone xenolith in picrite is surrounded by two reaction rims of picotite and olivine, and a xenolith of quartzite in tholeiite has a reaction rim of subcalcic ferroaugite.

The cases of pure melting of sediments described seem due to their nonrefractory character rather than to high temperature of the magma. The fused portion represents alkali feldspar and quartz, but quartz has been only partially melted for rounded grains still remain scattered through the glassy matrix.

MECHANISM OF INTRUSION

GENERAL STATEMENT

The close connection between the Karroo dolerites and the Stormberg lavas has already been emphasized, but any account of the mechanism of magmatic ascent must consider other igneous provinces of Trias-Lias age connected with the folding of the Samfrau Geosyncline. Brief mention is therefore made of igneous activity in African areas outside the Karroo Basin and also in South America, Antarctica, and Tasmania.

Much of this section is necessarily speculative and is written from the point of view that Continental Drift was a prime factor at this period.

DOLERITES AND LAVAS WITHIN THE KARROO BASIN

The field relations of the Karroo dolerites and Stormberg basalts have been discussed. A focus of igneous activity was noted in the Basutoland block and the area immediately southeast, and the authors believe that most of the intrusions in the Karroo Basin were formed by lateral spreading of the magma from this focus. In this focal region are found the sunken areas of Maclear and Barkly East, the thick sheets of Insizwa, Ingeli, and Mount Prospect, and the bell-jar intrusions of Mount Arthur and Bird's River Siding. The area is also characterized by a dike swarm trending northwest, with a weaker set at right angles. The sediments are riddled with dolerite; the ratio of intrusive to sedimentary material is estimated at 20 per cent (Du Toit, 1920, p. 10). Finally, the Kokstad type of dolerite is best developed in this focal area (Table 1). The position of the focal region, within the syntaxis of the Southern Folded Range and the Natal-Lebombo monocline, is also significant.

The age relations of the dolerite intrusions are not fully known, but Blignaut and Furter (1940, p. 72) distinguished 13 periods of intrusions in Northern Natal. Both at New Amalfi and Mount Arthur, 3 intrusive periods were recognized (Poldervaart, 1944; 1946).

Generally, the thinner dikes appear to be younger than the sills, while some of he larger dikes (*e.g.*, those of Fishback and Elephant's Head) acted as feeders to the sills.

More definite information is available with regard to the relative ages of the effusive and intrusive phases. Both dikes and sheets of dolerite are found in the Drakensberg basalts, while in the Stormberg region and in East Griqualand dikes of dolerite become vesicular as they rise through the uppermost Karroo strata (Du Toit, 1910a; 1911b). Hence the Stormberg vulcanicity must have preceded the main intrusive phase, with probably a certain overlap in time.

Though most of the Drakensberg lavas were erupted from fissures, now indicated by dolerite dikes, many shallow-rooted vents of central type occur in the Basutoland tract. The great majority originated through gaseous explosions, and no large quantities of basalt were ever erupted from them. Outside the Basutoland area, a few pipes, presumably of Karroo age, are found in the Carnarvon district and the Zuurberg. Here also the pipe fillings indicate shallow-rooted diatremes.

INTRUSION OF THE KAROO DOLERITES

Du Toit (1920, p. 28–36) suggested that the order of injection of the Karroo dolerites was from the summit downward toward the base. After the eruption of the Stormberg basalts, chiefly from fissures,

> "... the loading of the region by the immense pile of erupted matter, many thousands of feet in thickness, prevented any further relief of intratelluric pressure in this manner; hence, impelled from below, the magma burst up into the cake of fissile strata supporting the basaltic covering, making its way along narrow conduits towards the uppermost sedimentary zones. Strengthened as a result of the consolidation of the newly acquired intrusive matter, this topmost zone now became much more resistant to injection, so that in turn lower and lower horizons were progressively invaded, the magma spreading laterally for the most part" (Du Toit, 1920, p. 33).

The present authors confirmed this hypothesis in a number of cases, the most striking of which are at Elephant's Head and New Amalfi. The Elephant's Head dike was followed by Stockley (personal communication) in Basutoland over 37 miles. The dike is 300 yards wide, vertical, and forms a continuous exposure in Basutoland. In the Transkei it was followed eastward for 16 miles, from the Basutoland border until it spread out along a thick argillaceous horizon of the Burghersdorp beds, as the New Amalfi sheet. In this area its course is sinuous, the intrusion becomes roofed and probably floored, and in a series of intermittent exposures the dike descends the Drakensberg over a vertical distance of 3000 feet. While its field relations in Basutoland do not exclude the possibility that it acted as a fissure through which the basaltic lavas reached the surface, the relations in the Transkei indicate that the magma subsequently spread laterally and, reaching lower and lower levels, finally intruded as a thick sheet along the first major argillaceous horizon encountered.

The authors agree with Scholtz (1936, p. 203) that the Karroo dolerites were emplaced in a series of magmatic heaves following closely after one another. Earlier magmatic impulses were generally of undifferentiated olivine dolerite magma, but during the closing stages of igneous activity differentiated portions of the magma were intruded in small quantities, more acidic phases (Hangnest and Kentani type) as well as basic portions (Kranskop type and picrites). During the chief period of

igneous activity, relatively undifferentiated basalt magma was forced into and through the strata of the Karroo Basin, in marked contrast to the lavas and intrusives of the outlying areas.

Within the Karroo Basin the dolerites were generally intruded along planes of least resistance—*i.e.*, major argillaceous horizons in the Karroo strata. The magma also

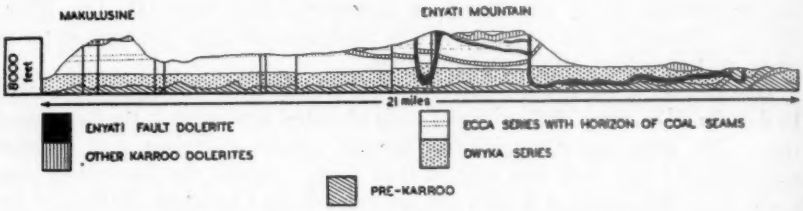


FIGURE 40.—*Section through the northern Natal Coalfield*
To illustrate intrusion of Karroo dolerite sills by uplift of the roof. (Modified from Blignaut and Furter, 1940).

worked its way along vertical planes of weakness, such as joint planes and shallow faults, pre-existent in the Karroo beds. Thus it was noted that sheets range from horizontal to vertical, in thickly bedded arenaceous strata (*e.g.*, Molteno beds) where "stepped" and "corrugated" sheets are common. The sinuous course of the Elephant's Head dike is believed to have been determined by such a combination of joint planes and shallow faults in the Molteno beds. Another example of an abnormal intrusion is the Enyati fault intrusion, which

> "... differs from the true dikes in that it has no regard for direction, but meanders in an unbelievable manner, and varies abruptly from a few inches to 300 feet. In some localities it is resolved into a dike and a sill, these always merging again some distance away" (Blignaut and Furter, 1940, p. 59).

The magma probably spread gradually through the Basin in a succession of heaves, ascending, descending, and ramifying freely, but always seeking planes of least resistance, either horizontal or vertical. High intrusive force is evidenced only in the straight and narrow dikes, injected after the major sill phase. The dominant form of the Karroo intrusions is the concordant and more or less horizontal sill, which may be emplaced by uplift of the roof or by sagging of the floor, or by a combination of both. Levinson-Lessing (1932, p. 281) believes that the former mode of emplacement is typical of subsiding regions, while the latter characterizes areas experiencing uplift. In the Karroo the chief mode of emplacement appears to have been through uplift of the cover, a feature particularly well illustrated in Northern Natal (Fig. 40). In the focal region of Basutoland and the Transkei, however, emplacement was effected both by uplift of the roof and sagging of the floor of the intrusions. Thus Scholtz (1936, p. 203) concludes that the Insizwa sheet was injected in three or more magmatic impulses, through sagging of the floor of the intrusion. In the same area sedimentary blocks, isolated by ring fractures also subsided, the dolerite magma occupying the potential vacua created above them, as at Mount Arthur.

Accepting du Toit's estimate for the focal region of a dolerite:sediment ratio of 1:4, over 4500 feet of dolerite is calculated to have been intruded into the Karroo

sediments. Possibly this ratio decreases to about 10 per cent for the outer areas of the Karroo Basin, and an estimate of 2000–3000 feet for the average uplift effected by the intrusion of the dolerites seems reasonable. The intrusion of both sills and dikes resulted in a lessening of tension produced by prolonged sedimentation within the Karroo Basin.

OUTLYING AREAS IN SOUTHERN AFRICA

Information concerning the Trias-Lias igneous activity in the outlying areas is scarce but permits a few general observations. Brief summaries and full references of the areas are given by du Toit (1939). They include the lavas of the Springbok Flats, the Lebombo, Limpopo, Tuli-Sabi River, Lupata Gorge, Victoria Falls, Gokwe, Bulawayo, Serokwe-Mashoro, Lake Makarikari, Lake Ngami, Mariental, Aminuis, Erongo, Brandberg, Messum, Cape Cross, and Kaokoveld.

Roughly, these may be divided into (1) Western area, including the Erongo, Brandberg, Messum, Cape Cross, and Kaokoveld complexes, (2) Central area, including the lavas of Mariental, Aminuis, Bechuanaland, Springbok Flats, and those north and northwest of Bulawayo, and (3) Eastern area, including the Lebombo, Limpopo, Tuli-Sabi River, and Lupata Gorge.

Western area. Highly differentiated lavas occur here, including basalts, porphyrites, and alkaline types, while the extrusive phase is followed by plutons of diorite, granophyre, and granite. According to Stahl (1932) the area shows block faulting and narrow graben structures, directed mainly northwest, but also northeast. The masses of the Erongo, Brandberg, Messum, and Cape Cross are thought to have originated through subsidence of isolated blocks with concomitant rising of differentiated magma, in several pulses, along the fault planes (Cloos, 1919).

Central area. The lavas of the Central area are essentially undifferentiated, amygdaloidal basalts which resemble the Basutoland basalts in petrography but differ from them chemically, particularly in their proportions of SiO_2, Al_2O_3, and MgO (lower), and TiO_2 and Fe_2O_3 (higher). In this respect they show closer resemblance to the normal basalts of the Western and Eastern areas.

Eastern area. As in the Western area the extrusives show strong differentiation. A succession basalt-rhyolite-basalt is recorded from the Lebombo (du Toit, 1929b) and the Lupata Gorge areas (Dixey and Campbell Smith, 1929). Limburgite and nepheline basalt occur at the base of the Lebombo in the Limpopo region (Rogers, 1925). Near Komati Poort a succession limburgite-basalt-rhyolite-basalt-alkaline lavas is found. The lower flows of limburgite disappear in the southern part of the Lebombo. Plutons of gabbro, augite granophyre, and microgranite occur throughout the Lebombo belt and in the Rhodesian sector occupy a large area. Alkaline plutons are found in the Lupata Gorge area.

The Lebombo belt, running due north for 660 kms, is situated on an easterly dipping monocline, which, south of latitude 28°, curves southwesterly to disappear into the sea near the Embotyi River mouth. Du Toit (1929b) showed that the monocline was formed during the eruption of the lavas, the main period of warping coinciding with the extrusion of rhyolite. The Lebombo dike swarm, directed due north, indicates that crustal tension in an east-west direction operated throughout the eruptive period.

TRIAS-LIAS VULCANITY IN GONDWANALAND

The mechanism of Trias-Lias vulcanicity can only be properly appreciated when seen as a phase in the history of greater Gondwanaland. Du Toit kindly supervised

a new reconstruction of Gondwanaland for the authors, including the latest available data. This reconstruction differs from du Toit's previous plan (1938, p. 58) in that Antarctica has been rotated slightly, bringing Australia nearer India and considerably shortening the Samfrau Geosyncline, which now passes through Antarctica (Fig. 41).

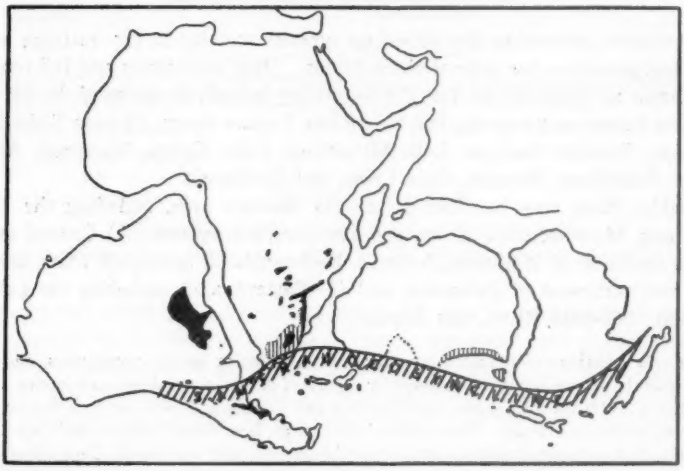


FIGURE 41.—*Reconstruction of Gondwanaland during Trias-Lias times*
(Modified from Du Toit, 1938).

South America. Two lava fields occur north of the Samfrau, that of the Parana Basin being the larger (Fig. 42) and occupying an oval area 1900 kms long. The extrusive rocks show considerable differentiation and resemble chemically the Kaokoveld lavas rather than those of Basutoland. The lavas of the Mendoza district, Argentina, include basalts, rhyolites, and porphyrites and are intruded by pre-Liassic plutons of granite and quartz monzonite. The extrusive Trias-Lias lava fields of Neuquen and Patagonia occur south of the Samfrau. The highly differentiated lavas include basalts, rhyolites, keratophyres, and tuffs, while in places they overlie the earlier-erupted porphyrites.
Falkland Islands. Besides the Zuurberg lavas, the Falkland Island dolerites are the only other Trias-Lias igneous rocks within the Samfrau. However, the islands may have formed part of the "back deep" of the Samfrau. No lavas are represented here, but undifferentiated dolerite, mainly olivine dolerite with rare bronzite dolerite, marks the intrusive phase of igneous activity (Baker, 1922).
Antarctica and Antarctic Archipelago. Dissappointingly little is known of the vital area of Antarctica, particularly its western part (Hearstland and Coates Land), facing East Africa and the Antarctic Archipelago. The intrusive phase of Trias-Lias igneous activity is represented by dolerite sills extending from Adelie Land to the Queen Maud Ranges. These closely resemble the intrusive bronzite dolerites of Tasmania (Browne, 1923).
Tasmania. As shown by Edwards (1942), the Trias-Lias igneous phase in Tasmania is represented by dikes and sills of bronzite dolerite, resembling the dolerites of Adelie Land and generally slightly more acid than the Karroo dolerites.
Australia and New Zealand. Trias-Lias igneous activity is represented in New South Wales (Gun-

nedah) and Queensland (Brisbane and Esk series), while the Bundanba series were invaded by large masses of granodiorite and syenite. In New Zealand, rhyolites and tuffs were erupted during the same period in the Coromandel and Canterbury Plains.

KARROO IGNEOUS PHASE IN RELATION TO GONDWANALAND

Three tectonic features dominated magmatic ascent in Southern Africa during Trias-Lias times—the Samfrau Geosyncline, the Karroo Basin, and the Natal-Lebombo Monocline (Fig. 42).

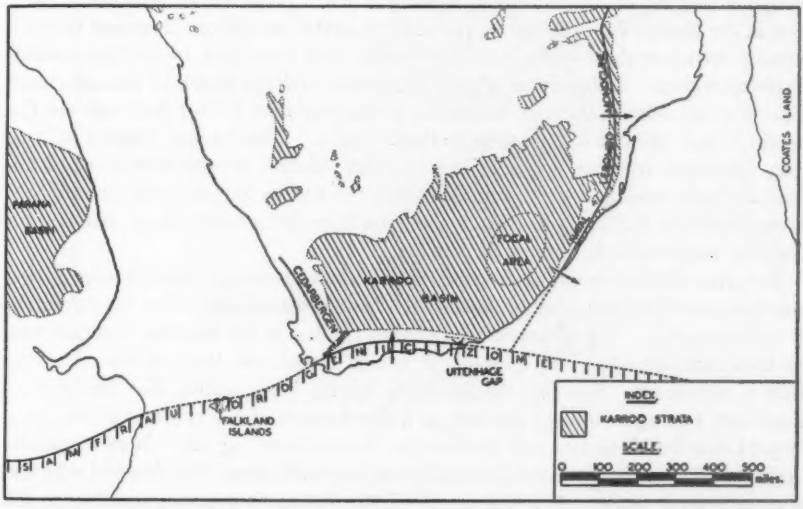


FIGURE 42.—*Relationship of Africa and Antarctica at the close of the Karroo period*

Prolonged sedimentation in the Samfrau Geosyncline terminated in its final collapse, producing the Cape Folded Belt in the extreme south of Africa. Four main periods of flexuring have been recognized, commencing in late Witteberg time and reaching a maximum intensity during early Molteno time.

The Samfrau Orogenic Belt was flanked to the north and south by troughs of subsidence, the foredeep to the north and the backdeep to the south, assuming the pressure to have come from the south, as indicated by northerly overfolding in the Cape Folded Belt. In Southern Africa the Karroo Basin was bounded to the west by the open foldings of the Cedarbergen, which rose together with the Southern Folded Belt, and were superimposed on the earlier Nama foldings. Deposition in this basin kept pace with subsidence over a prolonged period, producing a vast thickness of sediments. Other basins of deposition were formed in the foredeep and backdeep (*e.g.*, Parana Basin, Congo Basin). Finally, tension in the basins of the two deeps exceeded the strength of the crust, and magmatic wedges were driven into the sial, reaching the surface through tension clefts and spreading out as vast lava fields. Volcanic activity was followed by an even more extensive intrusive phase, during

which sills and dikes pierced the sedimentary cake beneath the lava. The bonds of Gondwanaland were loosened, and the shape of the present continents was formed during this period. Fissures in the Parana Basin and the Kaokoveld, acting as conduits for the magma, are directed northwest, while block faulting and graben structures, associated with the cauldron subsidences of the Erongo and Brandberg, may mark the beginning of the Africa-South America rift. Of greater importance in the present connection is the Natal-Lebombo Monocline formed during this time. The Lebombo lavas were erupted along this structure; their source was an elongated wedge of magma in which differentiation had taken place before extrusion. Tension in the Karroo Basin led to the pouring out of the Basutoland lavas and the subsequent intrusive phase of the Karroo dolerites, with focal area in the Basutoland-Transkei region. The position of the focal region and the trends of the associated dike swarms, when considered in relation to the east-west Folded Belt and the Lebombo-Natal Monocline, becomes entirely logical. The greater tension of this area, combined with the apparent lack of differentiation of both extrusive and intrusive phase, renders it highly probable that the Karroo fissures were directly connected with the basaltic substratum and that here the primary basalt magma underlying Gondwanaland rose to the surface.

The great rift that eventually separated Africa from Antarctica was initiated during this time, and the sea probably invaded the trough immediately after the formation of the monocline. The disappearance of that portion of the Samfrau together with its backdeep just where the north-south trough intersected the east-west Orogenic Belt is significant. Possibly the Zuurberg basalt, lying within the downfaulted Uitenhage gap, open toward the east, is a direct result of the syntaxis of the great Natal-Lebombo Monocline and the Samfrau Geosyncline (Fig. 42). If so, it appears quite logical that these basalts are undifferentiated and chemically identical with the Karroo igneous phase. Although a mid-Cretaceous age is generally ascribed to the east-west faults in the Southern Folded Belt, the authors consider that they were probably initiated in Trias-Lias times. The same applies to the northeast-trending faults in Natal and Zululand.

WEATHERING OF KARROO DOLERITES

MACROSCOPIC FEATURES

Karroo dolerites show great diversity in the degree and mode of weathering, depending on (1) climate, (2) local topography, (3) flow of ground water, (4) grain size, (5) chemical composition of the dolerite, and (6) type of jointing in the intrusions.

Although climatic conditions vary from subarid in the Western Karroo to subtropical in Natal and Zululand (Fig. 43), the influence of climate is more discernible in the soils derived from the dolerites than in the weathering of the rocks themselves. Dolerites are particularly susceptible to the action of water. Thus a favorable drainage pattern may produce pronounced weathering of the dolerite, even in the more arid parts of the Karroo. In general dolerites on escarpments and steep slopes are less deeply weathered than those in closed valleys, or exposed on "vloers", where drainage is bad, and surface or ground water tends to accumulate. Local basins of

decomposition may be formed where ground water is dammed by dolerite intrusions, and weathering may extend deeply at such places (Enslin, 1943). The grain size and chemical composition of the dolerites are also important in accounting for the intensity of the weathering. Rocks in which hydrothermal alteration is pronounced,

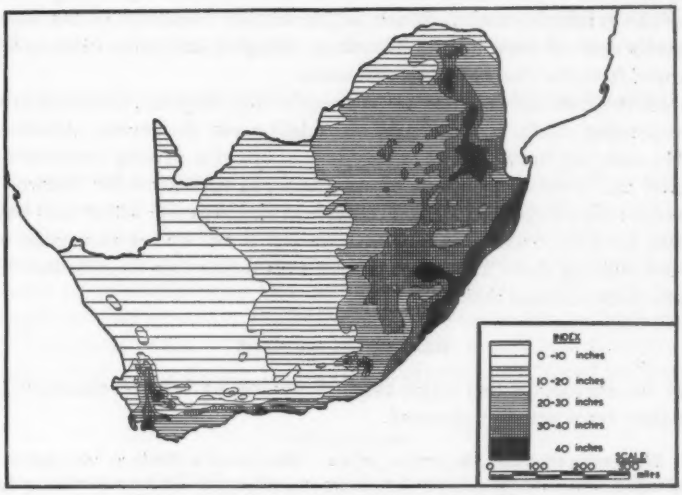


FIGURE 43.—*Isohyets for the Union of South Africa*
(Modified from Year Book for the Union of South Africa).

such as metasomatic granophyres, are prone to further atmospheric alteration. Fine-grained, ophitic dolerites of Perdekloof type are less subject to weathering than are coarser-grained dolerites. Dolerites of the Hangnest type break up very easily and weather even more readily than the associated Karroo siltstones (Pl. 6, fig. 1). Jointing aids downward circulation of water into the dolerite intrusions and helps to determine the type of outcrop produced and to govern, to a large extent, the shape of the broken joint blocks. The latter may be slender columns, angular, subangular or rounded blocks, or curved slabs. Columnar and angular or subangular blocks are always fresh and ring to the hammer, though covered with a thin weathering crust. The curved slabs and rounded blocks, on the other hand, generally have a thick weathering crust, while alteration frequently extends to the cores. In the more arid regions many of the joint blocks show a shiny black desert varnish. Spheroidal weathering and honeycomb weathering is common in Natal and the Orange Free State (King, 1942, Fig. 17, 18).

The weathered crust may be greenish brown or deep red. If the soil is very damp, the crust is often ocherous yellow, soft and spongy, though the dolerite core may be fresh (Beater, 1946). The deep-red crust is generally only 0.5 mm thick, while the core is fresh dolerite. The greenish-brown weathering crust usually forms friable exfoliating scales, which are often loose and may amount to a considerable thickness, while even the dolerite core may show signs of incipient alteration. In the Transkei

the red weathering crust may cover a bluish-green chloritic zone 1–2 mm thick, which in turn covers the fresh dolerite of the core. The two zones of alteration are sharply defined in the hand specimen, and the chloritic zone gives the impression of actually being fresher than the unaltered dolerite.

White incrustations of surface limestone are common and may occupy the spaces between the exfoliating slabs. Much of the surface limestone of the Karroo is undoubtedly derived from dolerite intrusions, though a large proportion must have been formed from the lime in Karroo sediments.

Soils derived from the dolerites are greenish brown, deep red, chocolate brown or black, depending chiefly on the degree of oxidation and the amount of humus present. The soils are heavy loams or clay loams, carrying varying amounts of sand and gravel and occasional nodules of limestone and clay (Van der Merwe, 1941). The dolerite soils are more fertile and retain heat and moisture better than those derived from the Karroo sediments. The latter fact is well known among the natives who prefer building their "kraals" on dolerite soils rather than on soils derived from the Karroo siltstones and shales.

MICROSCOPIC FEATURES

Under the microscope two major types of weathering become apparent, though intermediate types are also observed.

Type A. Weathering crust thin; deep red or yellow. Weathering is chiefly at the expense of the ferromagnesians, which alter to red or yellow-brown serpentine, with further alteration to hydrated iron oxides in more advanced stages of weathering. The alteration produces a considerable increase in volume as indicated by cracks (filled with serpentine) in the feldspar laths. Plagioclase, though fractured, remains clear and unaltered but in the outermost zones of the weathering crust it cannot be recognized. The crust is a tough mass of serpentine containing small grains of magnetite and varying amounts of hydrated iron oxides. The marked ability of the serpentine mantle to gain or lose water gives the weathering crust its power to protect the dolerite core from alteration.

Type B. Weathering crust thick; dirty brown or gray. Alteration has affected both feldspar and pyroxenes. Feldspar becomes turbid, and felty aggregates of clay material are developed. The pyroxenes develop basal striations, along which alteration to brown or green serpentine proceeds. The simultaneous alteration of plagioclase and pyroxene produces a friable, porous crust, which offers little protection to the dolerite core. Increases in volume resulting from the alteration of plagioclase and pyroxene are relieved by exfoliation, which also exposes new portions of the dolerite to the agents of weathering.

The alteration of pyroxenes was studied in more detail in weathering of type B, and the stages recognized are:

(1) Hydration and leaching of $FeSiO_3$. The pyroxene becomes progressively paler until colorless, at the same time acquiring a peculiar dusty appearance due to innumerable minute granules of hydrated iron oxide. The optic axial angle increases to values between 50° and 60°, an important fact, since many of the "diopside lamellae" observed in augite crystals (De Swardt and Murray, 1944, p. 8) may well be due to incipient alteration.

(2) Further hydration and removal of $CaSiO_3$. During the first stage strain in the crystal structure was relieved by the development of basal partings. In the second stage the crystal structure has collapsed. Extinction is ill-defined, and the "pyrox-

ene" areas are crowded with inclusions. Serpentinization, starting along basal partings during the first stage, has replaced nearly all the pyroxene in the second stage, leaving only a few small areas unaffected.

In a few instances the first stage—*i.e.*, leaching of iron—was also observed in olivines, just prior to their alteration to brown or green serpentine. The mineral becomes leaden gray, due to the development of numerous minute spots of opaque ore. A corresponding change in the optic axial angles was also noticed.

APPENDIX—DRAKENSBERG, ZUURBERG, AND LEBOMBO LAVAS

INTRODUCTION

Although a detailed treatise of the Trias-Lias vulcanicity in Southern Africa falls outside the scope of this paper, the intimate connection between the extrusive and intrusive phase of Karroo igneous activity calls for a brief discussion of the Karroo volcanics. Their distribution and field characters have been dealt with, also the mechanism of magmatic ascent and their relation with the Karroo dolerites. The following brief notes summarize to a large extent the work of other authors, in particular Du Toit (1904; 1910a, 1911b; 1920; 1929b; 1939), Haughton and Rogers (1924), Henderson (1909), Prior (1910), Rogers (1925), Schwarz (1902), and Stockley (1940; 1947).

PETROGRAPHY

Drakensberg lavas.—The Drakensberg lavas consist almost exclusively of basalt, petrographically very similar to the Karroo dolerites. Many of the lavas of the thicker flows are often quite coarse-grained and can be distinguished from Karroo dolerites only by the occurrence of small amygdules in their upper and lower portions. Alteration may be pronounced, particularly in those lavas in which gas cavities are most prominent. The latter are usually filled with quartz, agate, chalcedony, and calcite, or with zeolites such as heulandite, stilbite, thomsonite, and scolecite.

The lavas may be classified as: (1) olivine-rich basalts, (2) olivine-poor basalts, and (3) bronzite basalts, corresponding to respectively the Kokstad type, the Blaauwkrans and Perdekloof type, and the Hangnest type of Karroo dolerite. Some varieties are quite rich in olivine—*e.g.*, a coarse-grained basalt on the Dordrecht-Barkly East road which contains 8.5 per cent of olivine. Stockley (1940, p. 457) recorded two oceanites and one trachyte in a collection of 52 Stormberg basalts from Basutoland.

The plagioclase of all the basalts examined by the authors lies within the range labradorite-bytownite, and hence none could be classified as andesite according to modern terminology. (Cf. Du Toit, 1911b.)

Zuurberg lavas.—Although a few Zuurberg basalts contain analcite (Haughton and Rogers, 1924, p. 245), the bulk of the Zuurberg lavas closely resemble the Drakensberg volcanics. The analyzed rock (Table 28, no. 1) is a subophitic olivine basalt with amygdules filled with chlorite and analcite.

Lebombo lavas.—A greater variety of rock types is encountered in the Lebombo

volcanic belt, indicating differentiation in a deep-seated magmatic wedge, prior to extrusion (Du Toit, 1929b). In the Swaziland-Zululand portion of the Lebombo there is a succession basalt-rhyolite-basalt. Near Komati Poort this succession is further complicated by the occurrence of limburgite at the base of the sequence and of alkaline lavas toward the top. Yet further north alkaline lavas are also found at the base (Rogers, 1925).

The basalts of the southern Lebombo resemble those of the Drakensberg, but toward the north the Lebombo basalts show greater affinity to the volcanics of Southern Rhodesia, particularly with regard to their titania content. Thus many of the basalts north of Komati Poort contain titanaugite, while sphene is a prominent accessory of the northern rhyolites. South of Komati Poort both basalts and rhyolites are generally poor in titania, and the latter show a close chemical correspondence to the acid veins of the Karroo.

The rhyolites are glassy, sometimes partly or wholly devitrified, and contain phenocrysts of orthoclase and anorthoclase. More basic varieties show phenocrysts of oligoclase or andesine and merge into trachytes. The limburgites are characterized by phenocrysts of olivine and augite, set in a dark, glassy base. In a few instances the groundmass is crowded with minute crystals of augite.

CHEMICAL DATA

Table 27 gives chemical analyses of 1 Zuurberg basalt, 4 Drakensberg basalts, 1 Lebombo limburgite, and 2 Lebombo rhyolites. The new analyses were done for the authors by Mr. F. Herdsman. Two incomplete analyses of Lebombo rhyolites quoted by Henderson (1909) were not included.

The high soda of the Zuurberg analysis is accounted for by the analcite in amygdules scattered through the rock. The analysis of the Lebombo limburgite is remarkable in that the norm shows very little olivine. Apparently olivine crystallized in excess of the stoichiometric proportion, while quartz and plagioclase were quenched in the dark glassy base. In accordance with this view most of the olivine phenocrysts show evidence of resorption.

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TABLE 27. - CHEMICAL ANALYSES, NORMS, AND MODES OF KARROO LAVAS

ZUURBERG, DRAKENSBERG, AND LEBOMBO LAVAS.

CHEMICAL ANALYSES.

	a	b	c	d	e	f	g	h
SiO_2	50·75	48·43	49·95	51·74	54·15	49·02	66·40	71·41
TiO_2	1·13	1·25	·82	·87	·70	3·39	·54	·01
Al_2O_3	13·80	14·62	14·47	15·62	15·48	10·03	12·91	13·94
Fe_2O_3	4·65	4·97	2·27	·62	·67	1·82	3·87	·43
FeO	6·20	5·81	8·23	9·46	9·04	10·31	2·56	·78
MnO	·10	·22	·16	·18	·18	·46	·09	·09
MgO	7·19	6·47	8·08	7·16	6·35	12·23	·83	·09
CaO	8·90	10·69	10·12	10·36	9·27	7·72	2·32	·62
Na_2O	2·35	1·64	1·89	1·74	1·63	1·59	3·30	3·75
K_2O	·85	·64	1·47	·41	1·22	1·28	3·09	4·74
H_2O+	2·35	1·10	·76	·98	·83	1·23	3·16	4·09
H_2O-	1·85	3·90	·11	·35	·68	·48	1·09	·34
P_2O_5	·20	·20	·24	·03	·06	·16	·16	tr.
CO_2	nil	·01	nil	·64	·32	nil	nil	nil
TOTAL	100·78	100·15	100·07	100·18	100·22	99·89	100·02	100·49

NORMS.

Qu.	8·2	7·3	–	4·8	7·0	–	29·6	29·8
Or.	5·5	3·8	8·9	2·2	7·9	7·8	18·4	27·8
Ab.	24·1	13·6	16·2	14·7	13·8	13·6	27·8	32·0
An.	21·9	30·9	32·0	33·6	31·0	16·1	10·6	3·1
Cor.	–	–	–	–	–	–	·2	1·4
Di Wo.	8·9	8·9	7·8	8·7	3·2	8·9	–	–
Di En.	6·2	6·2	4·8	2·9	2·3	8·8	–	–
Di Fs.	2·0	2·0	2·6	2·6	2·6	2·8	–	–
Ol Fo	–	–	4·6	–	–	2·4	–	–
Ol Fa	–	–	3·1	–	–	·8	–	–
Hy En.	11·6	10·0	9·8	15·0	13·6	21·4	1·5	·2
Hy Of.	3·7	2·9	5·5	13·2	12·6	9·5	·8	1·1
Il.	2·3	2·4	1·6	1·7	1·4	6·5	1·0	–
Mt.	6·7	7·2	3·3	·9	·9	2·6	5·6	·9
Ap.	·3	·3	·3	–	·3	·3	·3	–
$CaCO_3$	–	–	–	1·5	·7	–	–	–
H_2O	4·2	5·0	·9	1·3	1·2	1·7	4·2	4·4
TOTAL	100·6	100·0	100·1	100·1	100·5	99·9	100·0	100·4

INDEX.

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b. Thakabanna's Cattle Post (25°26'S, 28°39'E), Drakensberg lava (analysis Royal Mint, Pretoria). Stockley, 1930.

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e. Barkly East (30°4'S, 27°27'E), Drakensberg lava. New Analysis.

f. Letaba (23°55'S, 31°33'E), Lebombo limburgite. New Analysis.

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PLATE 7.—PHOTOMICROGRAPHS OF KARROO DOLERITE TYPES

Figure

1.—Perdekloof type dolerite, Glen Grey sheet (Queenstown). × nicols (× 13).

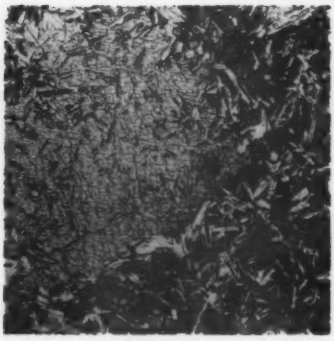
2.—Blaauwkrans type dolerite, Ventersburg (O.F.S.). Ordinary light (× 13). Note columnar, colorless pigeonite with heavy cross fractures.

3.—Kokstad type dolerite, New Amalfi (Matatiele). × nicols (× 13).

4.—Hangnest type dolerite, Hangnest (Calvinia). Ordinary light (× 13). Note conspicuous crystal of bronzite.

5.—Downes Mountain type dolerite, Hangnest (Calvinia). Ordinary light (× 13). Note curved pigeonite crystal.

6.—Hanover type dolerite, Hanover. × nicols (× 13). Note large bronzite crystal with thin mantle of augite.



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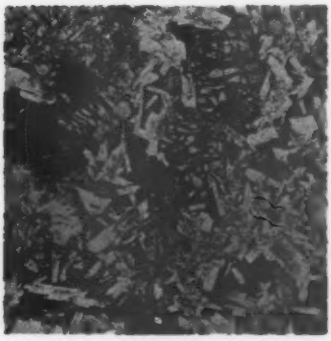

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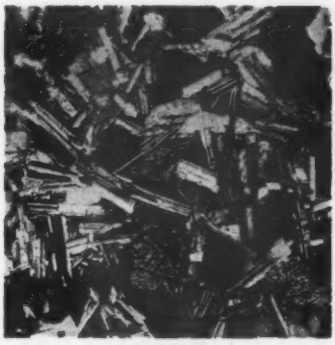

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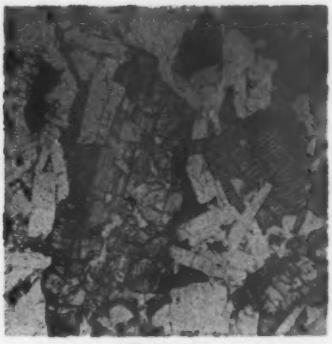

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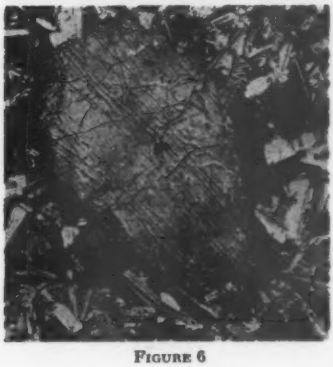
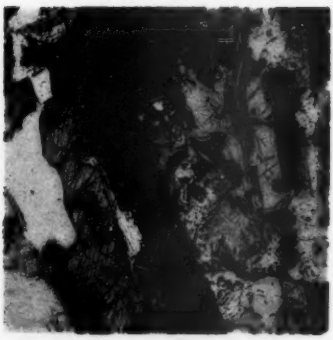


FIGURE 6

PHOTOMICROGRAPHS OF KARROO DOLERITE TYPES

FIGURE 1

FIGURE 2



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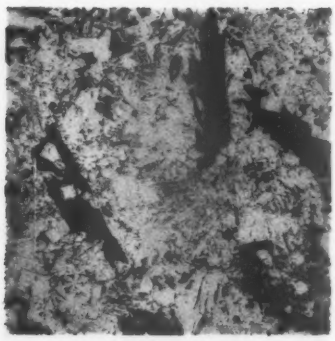

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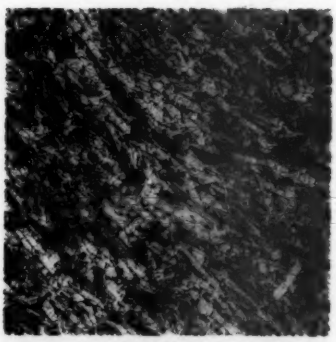


FIGURE 5

FIGURE 6

PHOTOMICROGRAPHS OF DOLERITE DIFFERENTIATES AND ASSOCIATED ROCKS

PLATE 8.—PHOTOMICROGRAPHS OF DOLERITE DIFFERENTIATES AND ASSOCIATED ROCKS

Figure

1.—Marginal tholeiite, Linton (Kokstad). Ordinary light (× 13).
2.—Picrite, Calamity Hill (Matatiele). × nicols (× 13).
3.—Dolerite pegmatite, Alewyn's Gat (Beaufort West). × nicols (× 13).
4.—Metasomatic granophyre, Hangnest (Calvinia). Ordinary light (× 13).
5.—Foliated picrite-dolerite, Umzimhlava Poort (Insizwa). × nicols (× 13).
6.—Contact basalt, Hangnest (Calvinia). Phenocrysts of plagioclase and bronzite. × nicols (× 13).

PLATE 9.—PHOTOMICROGRAPHS OF CONTACT BASALTS, SEDIMENTS AFFECTED BY DOLERITE, AND HYPERSTHENE WITH GRAPHIC INTERGROWTH

Figure

1.—Contact basalt, Theekloof (Fraserburg). Phenocrysts of plagioclase and olivine. Ordinary light (× 13).

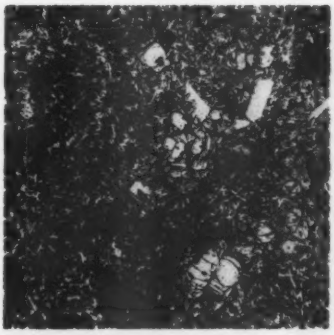
2.—Contact basalt, Blaauwkrans (Calvinia). Glomeroporphyritic plagioclase. × nicols (× 13).

3.—Glass of Ecca grit, Heilbron (O.F.S.). Ordinary light (× 13).

4.—Rheomorphic siltstone vein, Hangnest (Calvinia). Ordinary light (× 13).

5.—Rheomorphic breccia, Hangnest (Calvinia). Ordinary light (× 13).

6.—Graphic intergrowth in orthopyroxene, Palisade (New Jersey). × nicols (× 35).



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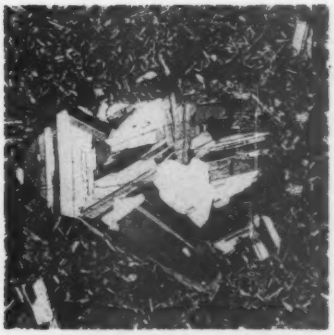

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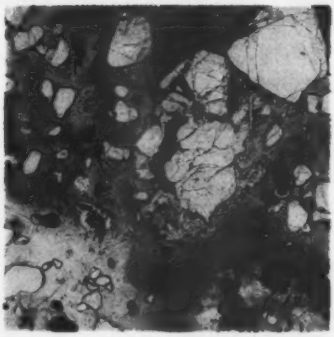

FIGURE 3

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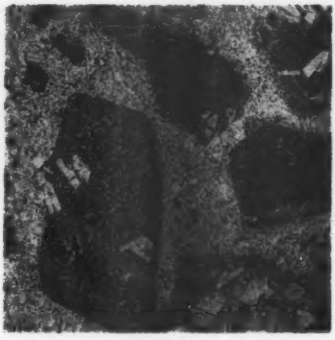

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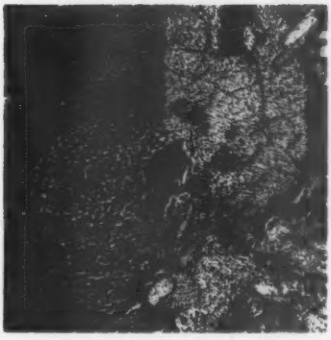


FIGURE 6

PHOTOMICROGRAPHS OF CONTACT BASALTS, SEDIMENTS AFFECTED BY DOLERITE, AND HYPERSTHENE WITH GRAPHIC INTERGROWTH

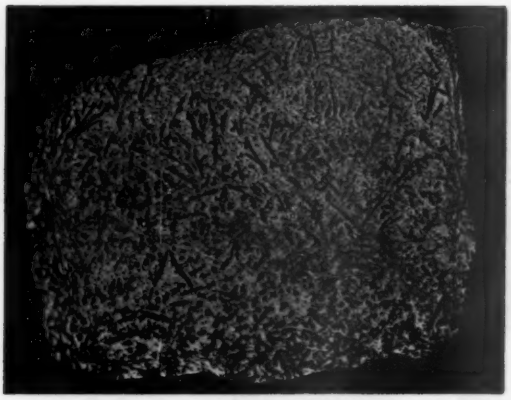


FIGURE 1

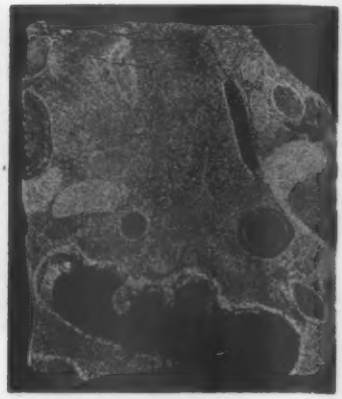


FIGURE 2

FIGURE 3

FIGURE 4

METASOMATIC AND ANATECTIC PHENOMENA IN THE KARROO

PLATE 10.—METASOMATIC AND ANATECTIC PHENOMENA IN THE KARROO

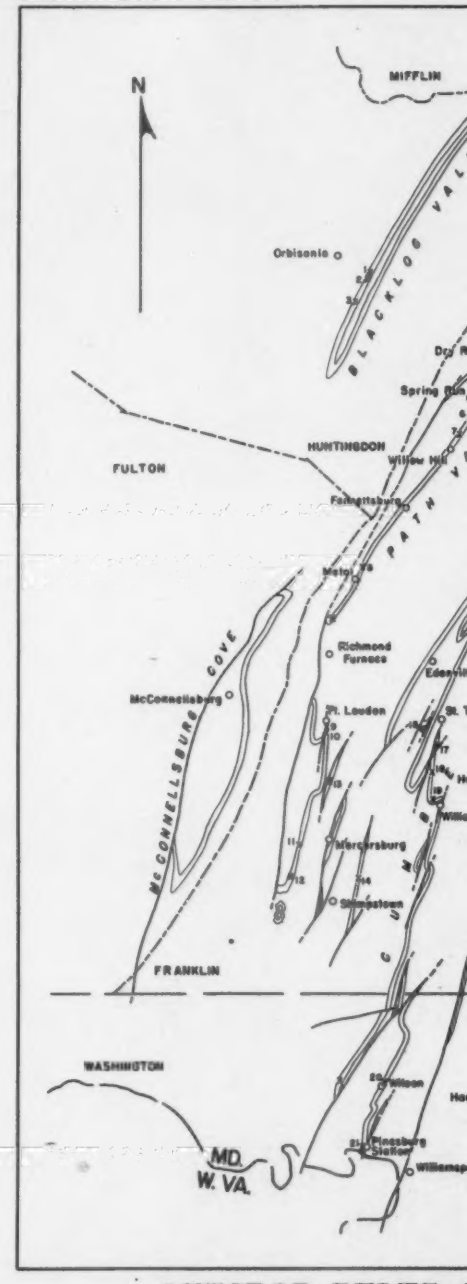
Figure

1.—Metasomatic granophyre, Alewyn's Gat (Beaufort West). × ½ natural size.

2.—Fine-grained granophyre from kernel zone, Rietkop (Calvinia). × ½ natural size.

3.—Fine-grained granophyre surrounding partly transfused Ecca siltstone, Rietkop (Calvinia). × ½ natural size.

4.—Rheomorphic vein of Beaufort siltstone in dolerite, Beaufort West reservoir. × ½ natural size.



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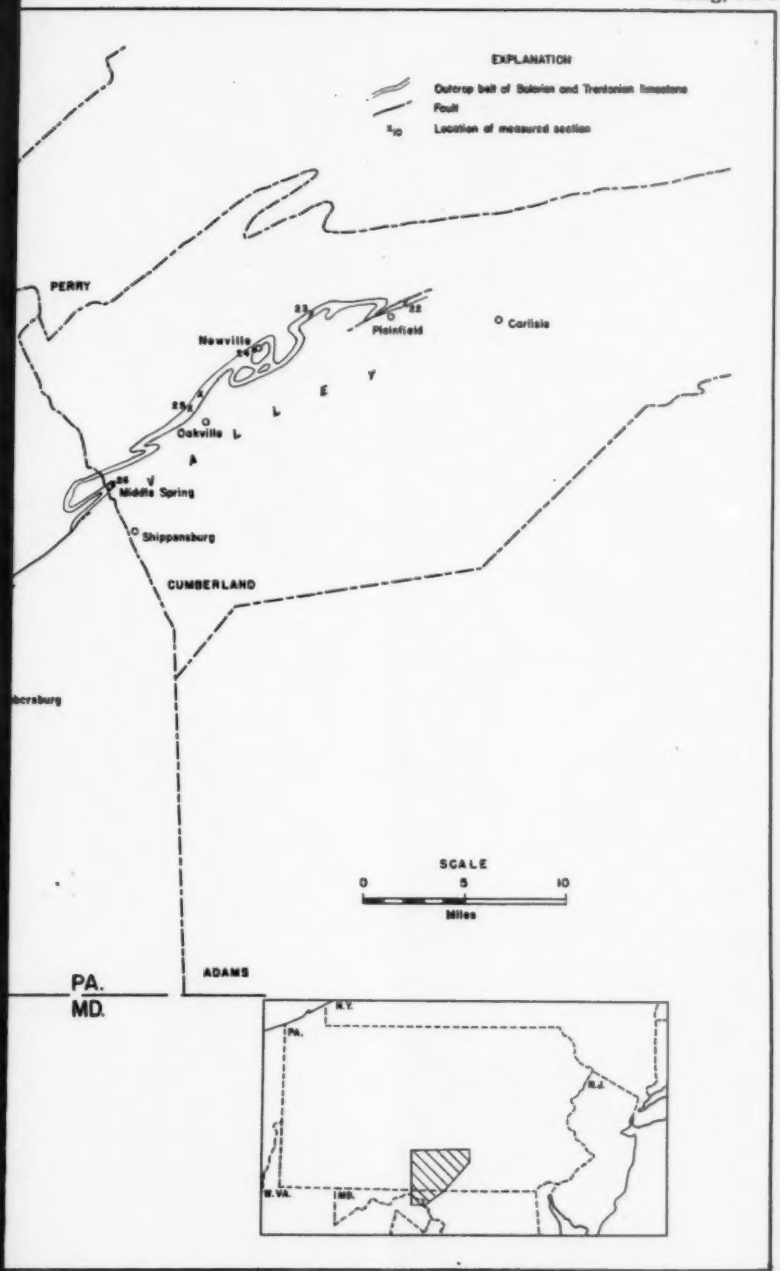

RIAN AND TRENTONIAN LIMESTONE
NTRAL PENNSYLVANIA

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LOWER MIDDLE ORDOVICIAN OF SOUTH-CENTRAL PENNSYLVANIA

BY LAWRENCE CAREY CRAIG

CONTENTS

	Page
Abstract	708
Introduction and acknowledgements	709
Regional setting	710
Historical review	711
Stratigraphic units	714
General discussion	714
"Stones River" limestone	715
Name	715
Description	717
Correlation	717
Shippensburg formation	717
Definition	717
Lithologic characteristics	718
Thickness	718
Pinesburg member of the Shippensburg formation	718
Definition	718
Lithologic characteristics	718
Thickness	719
Fossils	721
Correlation	722
Fannettsburg member of the Shippensburg formation	722
Definition	722
Lithologic characteristics	722
Thickness	725
Fossils	725
Correlation	726
Doylesburg member of the Shippensburg formation	727
Definition	727
Lithologic characteristics	727
Thickness	727
Fossils	729
Correlation	729
Mercersburg formation	731
Definition	731
Lithologic characteristics	731
Thickness	731
Housum member of the Mercersburg formation	732
Definition	732
Lithologic characteristics	732
Thickness	733
Fossils	733
Correlation	733

Page
Kauffman member of the Mercersburg formation 734
Definition 734
Lithologic characteristics 734
Thickness 735
Fossils 735
Correlation 738
Oranda formation 738
Name 738
Lithologic characteristics 739
Thickness 740
Fossils 740
Correlation 740
Martinsburg shale 742
Name 742
Lithologic characteristics 742
Thickness 744
Fossils 745
Correlation 745
Classification 746
Structural considerations 749
Paleogeographic interpretations 749
Summary 752
Geologic sections 753
References cited 778

ILLUSTRATIONS

Figure Page
1.—Correlation table of Middle Ordovician formations in central and south-central Pennsylvania 715
2.—Relation of Bolarian and Trentonian formations between western and eastern belts of outcrop, south-central Pensylvania 716
3.—Correlation of Salona metabentonites in south-central Pennsylvania 743
Plate Facing page
1.—Outcrop belts of Bolarian and Trentonian limestone in south-central Pennsylvania 707
2.—Relation of Bolarian and Trentonian formations in the western outcrop belts, south-central Pennsylvania 714
3.—Isopach map of the Shippensburg formation, south-central Pennsylvania 718
4.—Isopach map of the Mercersburg formation, south-central Pennsylvania 732
5.—Isopach map of the Oranda formation, south-central Pennsylvania 740
6.—Restored sections of Bolarian and Trentonian formations in south-central Pennsylvania .. 750

ABSTRACT

Broad paleostructural development in south-central Pennsylvania and Maryland was similar to that of central and eastern New York during the deposition of lower Middle Ordovician beds. Thick marine limestones of the easternmost belts of outcrop in the Cumberland Valley indicate the southward extension of the Champlain miogeosyncline. Thin and, in part, coarse-textured limestones in belts of outcrop just west of the Cumberland Valley indicate the extension of the Adirondack arch. Intermediate correlation shows the development of a relatively shallow platform east of the Adirondack arch and marginal to the Champlain miogeosyncline during part of the Bolarian (Kay, 1947) and Trentonian epochs.

Detailed description, definition and correlation of the stratigraphic units are presented. The Pinesburg and Fannettsburg members (new) of the Shippensburg

formation (redefined), the Housum and Kauffman members (new) of the Mercersburg formation (redefined), the Oranda formation (Cooper and Cooper, 1946) and the lower Martinsburg exhibit westward thinning by overlap or convergence or both from Champlain miogeosynclinal sections in eastern belts to sections on the Adirondack line in Path Valley. The Doylesburg member (new) of the Shippensburg is believed truncated to extinction southeastward from Path Valley by pre-Mercersburg regional unconformity. Argillaceous material in southeasternmost exposures of Oranda begins the Middle Ordovician transgressive clastic deposits resulting from the Vermontian disturbance.

Correlations indicate Shippensburg is post-middle Chazyan to pre-Chaumont in age, Mercersburg is Rockland and Kirkfield, Oranda is Shoreham, and lower Martinsburg is Denmark.

INTRODUCTION AND ACKNOWLEDGMENTS

Problems of correlation between the Chambersburg formation of the Appalachian Valley and beds of similar stratigraphic position near the Allegheny Front were first brought to the writer's attention in 1940. Although the relations of Middle Ordovician rocks across the Adirondack line are well known from the nearly continuous outcrop along the Mohawk Valley in New York, similar correlations to the south have been hampered by the absence of intermediate sections between east and west. Except for the central and south-central Pennsylvania area, intermediate sections are scarce southward to the latitude of Lexington, Virginia. Southward from Lexington, Appalachian thrusting progressively covers the intermediate sections between the eastern (allochthonous) and western (autochthonous) belts (Kay, 1941a, p. 1912; Prouty, 1946, p. 1189).

Folding, reverse faulting, and subsequent erosion have produced numerous belts of lower Middle Ordovician outcrop between Nittany Valley in central Pennsylvania and Chambersburg in south-central Pennsylvania. Marshall Kay (1941b, 1943a, 1944) has studied the belts of outcrop in central Pennsylvania. The study of outcrop belts in south-central Pennsylvania was undertaken to clarify correlation, determine the paleostructure and paleogeography, and draw comparisons with the Middle Ordovician of New York. Field work was begun in the summer of 1941 and preliminary conclusions were drawn (Craig, 1941b). Additional field work was done in the fall and winter, 1946–1947.

The correlations made in this study demonstrate the existence of both the Champlain miogeosyncline and the Adirondack arch as prominent structural features during the lower Middle Ordovician. Thus, the development of south-central Pennsylvania was broadly similar to that of New York during this time. However, a number of stratigraphic and structural details were different. During upper Bolarian and lower Trentonian deposition a shallow bench existed east of the Adirondack arch in south-central Pennsylvania. The Adirondack arch was a relatively positive structure in Pennsylvania from early Bolarian time, but in New York the Lower Trentonian, Rockland, beds are the first to show definitely its existence. Chaumont beds are present in the Champlain miogeosyncline of New York, but have not been recognized in south-central Pennsylvania and are thought to be absent due to nondeposition or subsequent erosion. An earlier inception of the Vermontian disturbance in Pennsylvania may be indicated by the influx of argillaceous material into the Champlain

miogeosyncline in Shoreham time. In New York the deposition of terrigenous material began in Denmark time.

The writer wishes to acknowledge his indebtedness to Professor Marshall Kay, of Columbia University, who suggested and supervised the study. Appreciation is expressed for suggestions made in the field by Dr. R. L. Miller, Prof. N. D. Newell, and Dr. C. E. Prouty. Thermal and X-ray analyses of several critical clay beds were made through the courtesy of Professor P. F. Kerr by Mr. H. D. Glass and Miss P. K. Hamilton, graduate students at Columbia University. A grant from the James Furman Kemp fund supported field work in 1946–1947.

REGIONAL SETTING

The area of study lies within the Ridge and Valley and Great Valley divisions of the Folded Appalachian province. The area is roughly triangular in outline and extends from Orbisonia and Carlisle, Pennsylvania, on the north, to Pinesburg Station, Washington County, Maryland, on the south (inset, Pl. 1).

The outcrop belts of lower Middle Ordovician may be grouped in six structural divisions: Blacklog Valley belt, McConnellsburg Cove belt, Path Valley belt, and western, central, and eastern belts of the Cumberland Valley. Blacklog Valley, McConnellsburg Cove, and Path Valley are anticlinal valleys on the northwest and west (Pl. 1). The Cumberland Valley exposes repeated belts of Middle Ordovician on the south and east. A faulted anticline and syncline in Franklin County, Pennsylvania, and Washington County, Maryland, divide these belts into three groups in the Cumberland Valley. The western belts are a series of fault slivers lying west of an anticline exposing "Stones River" limestone and older beds. These belts are exposed only in southwestern Franklin County, being mostly covered to the north and south by the younger sediments of North and Bear Pond Mountains. Central belts lie along the middle limb of the anticline-syncline from Pinesburg Station, Maryland, through St. Thomas, Pennsylvania, to near Upper Strasburg, Pennsylvania, where they join with the western belts in the plunging nose of the anticline. The adjacent syncline exposes Martinsburg shale and is the northern extension of the Massanutten syncline of Virginia. The eastern belts are a complex of folded and faulted belts of lower Middle Ordovician on the east limb of the Martinsburg syncline. They have been traced from the Pennsylvania-Maryland state line through Chambersburg to near Carlisle, but are faulted out for 4 miles between Chambersburg and Shippensburg.

Sections of Bolarian and Trentonian rocks have been measured in all the structural divisions. However, exposure of the outcrop belt of Bolarian and Trentonian rocks in McConnellsburg Cove is so poor that no satisfactory determination of stratigraphic sequence was made. The only earlier section from this valley found and measured in the course of this study has been discarded because of only partial exposure and indication of structural complication.

Authors have divided the Pennsylvania area into eastern belts and western belts due to the stratigraphic contrast delimited by the Martinsburg syncline. West of the syncline the interval between the "Stones River" limestone and the Martinsburg shale is thin and consists of a number of distinctive lithologic units. East of the syncline the interval is thicker and is not readily separable into lithologic divisions. The

use of "eastern belts" and "western belts" in this paper refers to this contrast. In Maryland all the sections studied are of the eastern type although they lie west of the Martinsburg syncline.

HISTORICAL REVIEW

G. W. Stose (1906, p. 211) defined the Chambersburg limestone as 1000+ feet of beds forming the top formation of the Shenandoah group, and stated that the upper part of the Chambersburg contained fossils which were referred to the Black River and Chazy by E. O. Ulrich. Later Stose (1908, p. 710–711) described four sections of the Chambersburg. One, 1½ miles west of Kauffman and 2.1 miles S.53°W. of Marion (section 27), is considered the type section, although it was referred to only as the most complete section in the belt of most typical development, the Chambersburg belt. The maximum thickness of the formation was given as 600 feet. Eight faunules were listed, four from eastern belts and four from western belts. The only correspondence recognized between the two groups of faunules was in the collections from the *Sinuites* zone, the fourth or upper faunule of each group. The following year Stose (1909, p. 8–10) amended the Chambersburg limestone in the Mercersburg-Chambersburg Folio. *Tetradium cellulosum*-bearing beds, 150 feet thick, were added to the bottom of the type section. The *Sinuites* zone was regarded as a transition bed to the overlying Martinsburg, but was retained in the Chambersburg. Ulrich (Stose, 1909) discussed five faunal zones in the eastern belts and five in the western belts. The five eastern faunules correspond to the four faunules presented in 1908 plus an additional faunule from the 150 feet of beds added to the base of the formation; from bottom to top they were referred to as (1) *Tetradium cellulosum*, (2) *Echinosphaerites*, (3) *Nidulites*, (4) *Christiania*, (5) *Sinuites*. The sequence of faunules from the western belts does not correspond with the earlier sequence (Table 1).

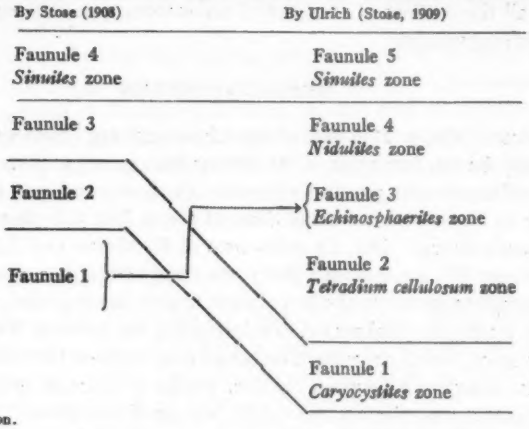
The most significant change is the apparent relocation of Faunule 1 of the earlier report as Faunule 3 in the later report. Except for three changes the two listings are identical: *Gonioceras chazyense* does not appear in Faunule 3; "*Cliocrinus* n. sp. ?" of Faunule 1 is represented by "*Cleiocrinus magnificus*" in Faunule 3; "*Echinosphaerites* sp. ? plates only" is represented by "*Echinosphaerites* sp." Ulrich correlated the *Echinosphaerites*, *Nidulites*, and *Sinuites* zones with the zones of the same names in the eastern belts. He regarded the *Caryocystites* zone as absent in eastern belts and the *Christiania* zone as absent in the western belts (Table 2).

Ulrich (1911, p. 328) subsequently restricted the Chambersburg limestone by referring the *Sinuites* zone to the Martinsburg shale.

Bassler (1919, p. 143–144) designated the Greencastle bed as a member at the top of the Chambersburg limestone and applied the term to a unit intervening between the *Christiania* and *Sinuites* zones and containing the *Christiania* zone faunule plus a recurrence of the genera, *Nidulites* and *Echinosphaerites*. The Greencastle bed was recognized only in the eastern belts of outcrop between the Pennsylvania-Maryland state line and Chambersburg.

Butts, Stose, and Jonas (1932, p. 16) restricted the Chambersburg limestone by referring the 150 feet of *Tetradium cellulosum*-bearing beds at the base to the Lowville limestone. The United States Geological Survey (Wilmarth, 1938, p. 395) accepted

TABLE 1.—*Comparison of faunule lists in western belts of Chambersburg limestone**



*Author's interpretation.

TABLE 2.—*Summary of E. O. Ulrich's correlations between eastern and western belts of Chambersburg limestone*

Western belts	Eastern belts
Faunule 5 *Sinuites* zone	Faunule 5 *Sinuites* zone
Absent	Faunule 4 *Christiania* zone
Faunule 4 *Nidulites* zone	Faunule 3 *Nidulites* zone
Faunule 3 *Echinosphaerites* zone	Faunule 2 *Echinosphaerite* zone
Faunule 2 *Tetradium cellulosum* zone	Faunule 1 *Tetradium cellulosum* zone
Faunule 1 *Caryocystites* zone	Absent

the formation at this stage of its emendation, regarding it as post-Lowville Black River in age.

The type section of the Chambersburg limestone was divided into three formations, Shippensburg, Mercersburg, and Greencastle (Craig, 1941b) as a result of fundamental disagreement with previous faunule zone correlations. Comparison of sections from eastern and western belts showed that (1) the cobbly "*Echinosphaerites*" zone (and lower part of the *Nidulites* zone) underlay the *Caryocystites*-bearing zone in

western belts; (2) the *Nidulites* zone contained a *Caryocystites*-bearing interval at the type section and occurred both above and below the *Caryocystites*-bearing and *Tetradium cellulosum*-bearing zones of western belts; (3) an upper *Echinosphaerites* zone was present locally in western belts, above rather than below the *Nidulites* beds, in a

TABLE 3.—*Correlation of Lower Middle Ordovician formations and generalized faunal zones in south-central Pennsylvania (Craig, 1941b)*

Western belts		Eastern belts	
Greencastle formation	*Sinuites* zone *"Christiania"* zone Upper *Echinosphaerites* zone	*Sinuites* zone *Christiania* zone Upper *Echinosphaerites* zone	Greencastle formation
Mercersburg formation	Upper *Nidulites*	Upper *Nidulites*	Mercersburg formation
Snyder member Benner formation	*Tetradium cellulosum* zone		
Hatter formation	*Caryocystites* zone	*Caryocystites* zone	
Shippensburg formation	Lower *Nidulites* zone Lower *"Echinosphaerites"* zone	Lower *Nidulites* zone Lower *Echinosphaerites* zone	Shippensburg formation
"Stones River" limestone	?	*Tetradium cellulosum* zone	"Stones River" limestone

position corresponding to the *Echinosphaerites* in the Greencastle bed at the type section (Table 3).

The absence in the eastern belts of distinctive lithologic characteristics led to the belief that the Hatter formation and the Snyder member of the Benner formation were not represented. However, Cooper and Cooper (1946, p. 56) have correlated the *Caryocystites* beds of St. Thomas (in the western belts) with the *Caryocystites*-bearing interval of the type Chambersburg section.

The beds included in the Shippensburg, Mercersburg, and Greencastle were identical with the beds Stose (1908, p. 710) included in the first description of the Chambersburg. Both bottom and top of the interval were the prominent lithologic contacts.

The original sequence of western belt faunules listed by Stose (1908, p. 713–714) is similar to that now recognized. The subsequent misplacement of the *Echinosphaerites* zone provided an incorrect correspondence between eastern and western

belt faunules and contributed to the example of oscillation presented by Ulrich (1911, p. 328–329).

STRATIGRAPHIC UNITS

GENERAL DISCUSSION

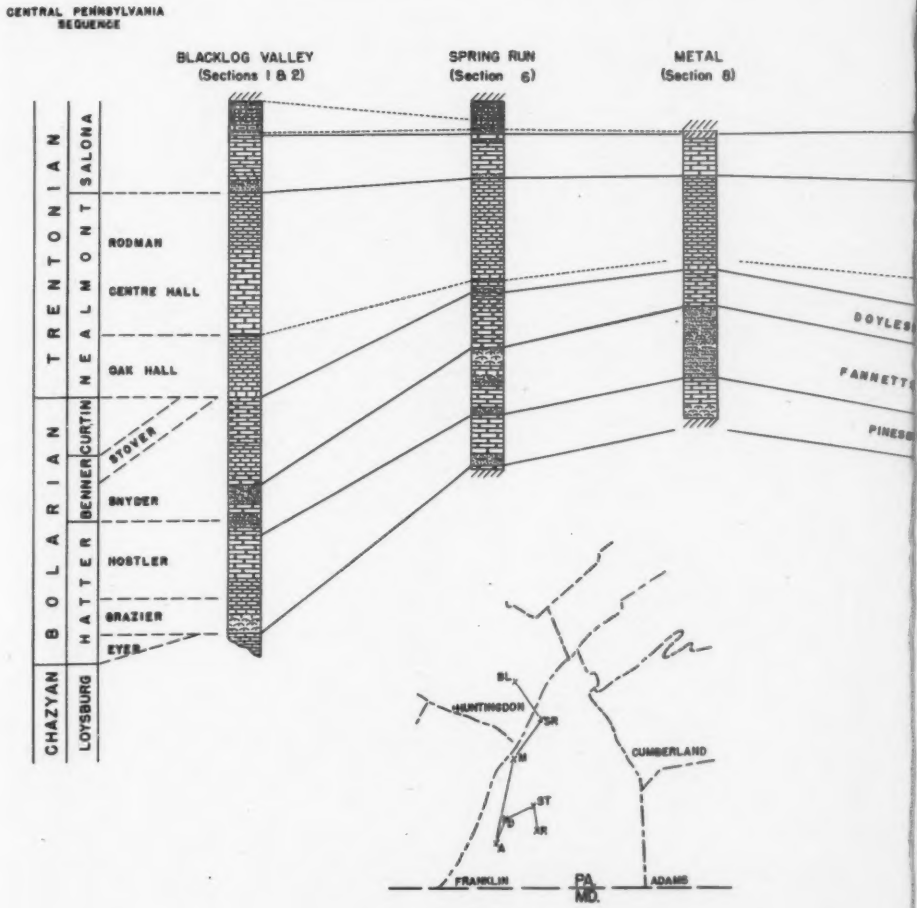
Additional work has shown the necessity for modifying the classification proposed in 1941.

Five time-stratigraphic units have been readily distinguished and correlated within the western belts of lower Middle Ordovician limestone in south-central Pennsylvania (Fig. 1; Pl. 2). Five units, not entirely the same, have been distinguished and correlated with greater difficulty in the eastern belts of outcrop (Fig. 1). General correlation between the eastern and western belts (Fig. 2) is regarded as certain, although some precision correlations are necessarily founded on meager evidence. The units of south-central Pennsylvania correlate directly with the central Pennsylvania sequence in Blacklog Valley.

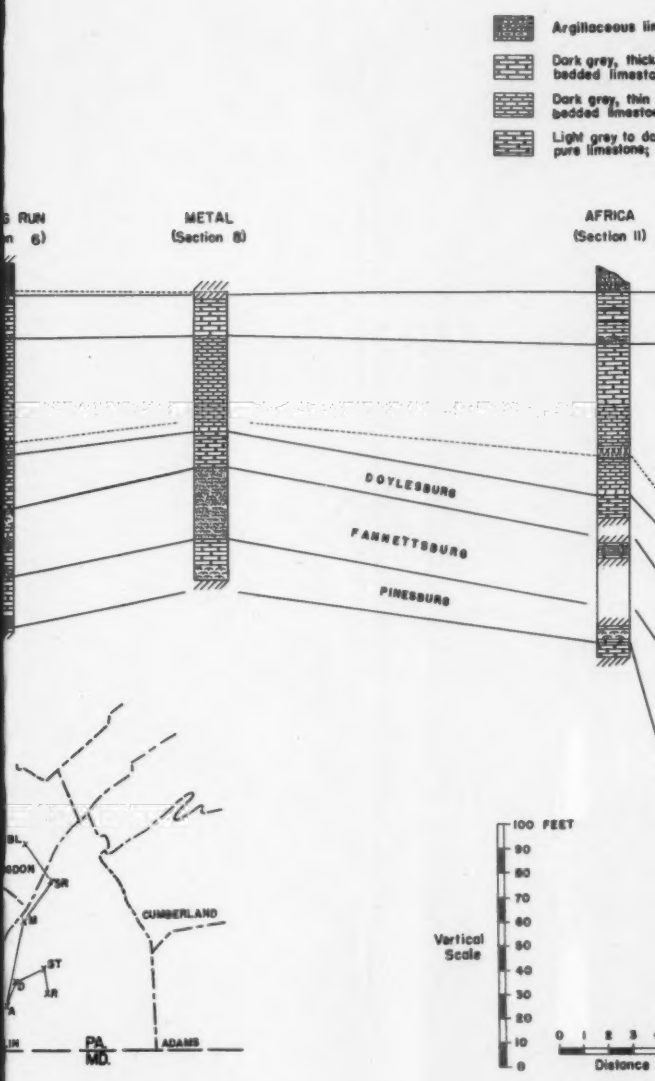
A number of distinctive yellow clay beds in the lower Middle Ordovician sequence have been used extensively in intersectional correlations. The clays are interpreted as beds of altered volcanic ash and, as such, denote essentially contemporaneous horizons where the same beds can be recognized in separate sections. Rosenkrans (1934) described the physical, chemical, and mineralogic properties of these clays; noted suggestions of volcanic structures in a few of his thin sections; and recorded the presence of allogenic biotite, zircon, apatite, and feldspar. No microscopic investigation of the clays was made during this study and no relic volcanic structures were observed. Biotite flakes are associated with several of the clays. A number of authors have referred to these eastern clays as bentonites, but more recently the term metabentonite (Ross, 1928) has found increased use. "Metabentonite" is used here for it indicates the contrast between these clays and the younger western bentonites by emphasizing the post-depositional alterations and implying the absence of commercially valuable properties in the eastern clays. The metabentonites neither absorb appreciable amounts of water nor disintegrate readily in water. They are more indurated than most bentonites and the published analyses indicate a higher potash content.

Kerr determined that these clays yield a characteristic diffraction pattern indicating an unnamed clay mineral different from those of the montmorillonite-beidellite and kaolinite groups. As a check of the clays encountered in this study, Mr. H. D. Glass and Miss P. K. Hamilton made X-ray and thermal analyses of eight clay samples taken from various clay beds at several locations in south-central Pennsylvania and Maryland. The majority of samples were collected from beds of unusual appearance or thinness. In addition, a sample from a bed of typical character was collected and analyzed as a control. All diffraction patterns gave gratifying comparisons with patterns of previously analyzed metabentonites. The thermal curves supported the assignment of the clays to the potash-bearing group.

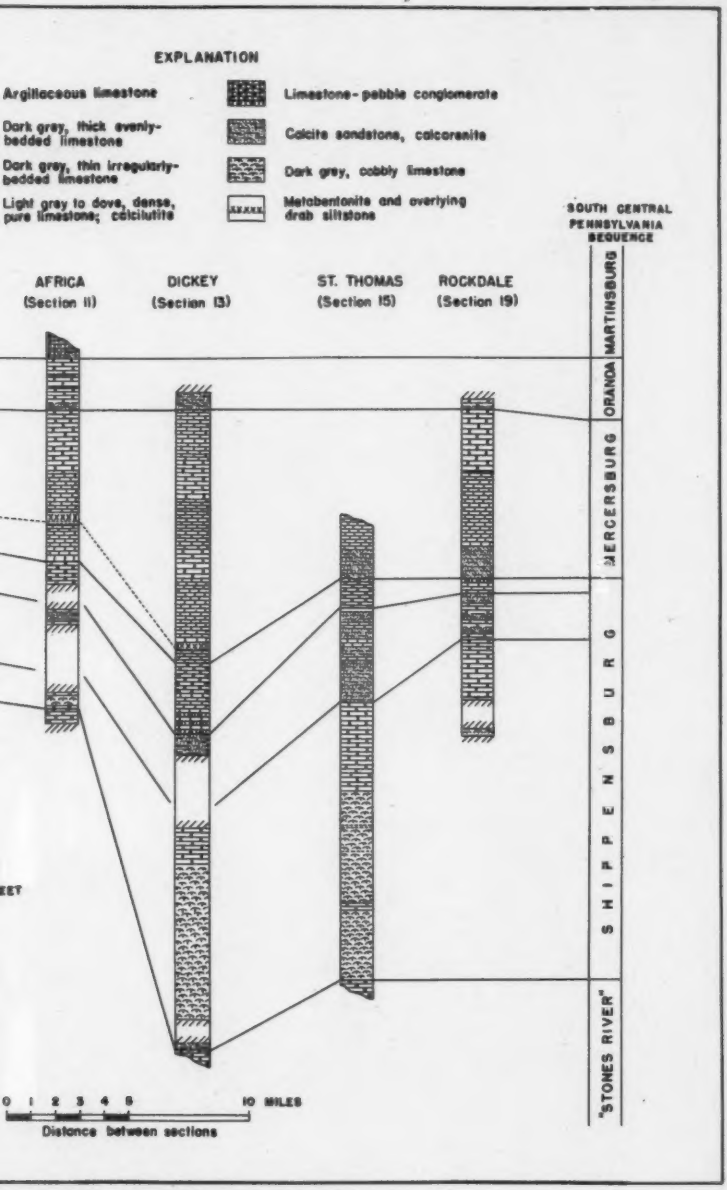
Most of the faunal zones of previous authors have been recognized. They are approximate, indistinct and useful only for general reference. The *Christiania* zone



RELATION OF BOLARIAN AND TREN
WESTERN OUTCROP BELTS, SOUTH



F BOLARIAN AND TRENTONIAN FORM
OUTCROP BELTS, SOUTH-CENTRAL PE



FORMATIONS IN THE AL PENNSYLVANIA

has not been identified. Neither a separate faunule nor a distinctive lithologic character has been found above beds referable to the *Nidulites* zone and below the upper *Echinosphaerites* occurrence.

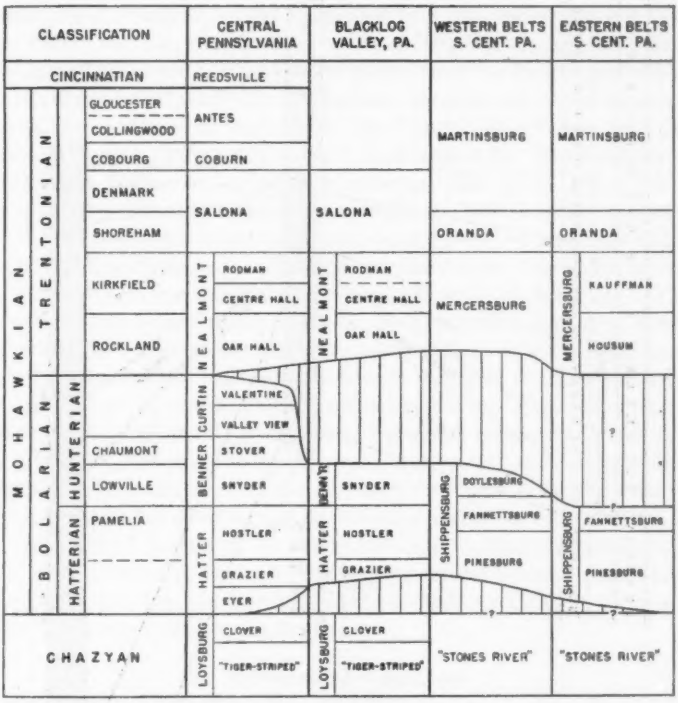


FIGURE 1.—*Correlation table of Middle Ordovician formations in central and south-central Pennsylvania*

Discrepancies in thickness determinations further complicate the recognition of faunal zones. Previously published thicknesses for the Chambersburg interval at the type section, 2 miles southwest of Marion, are about 100 feet less than that determined by the writer.

"STONES RIVER" LIMESTONE

Name.—The limestone beds overlying the Beekmantown and underlying the Chambersburg in Franklin County, Pennsylvania, were called Stones River by G. W. Stose (1906, p. 211; 1908, p. 707–709). B. N. Cooper (1945, p. 262–275) has shown that beds in the Appalachian region of Virginia have been consistently miscorrelated with the true Stones River of central Tennessee. Correlation from south-central Pennsylvania to Virginia indicates that the term, Stones River, was misused in the Chambersburg area and that equivalents of the true Stones River are probably within the

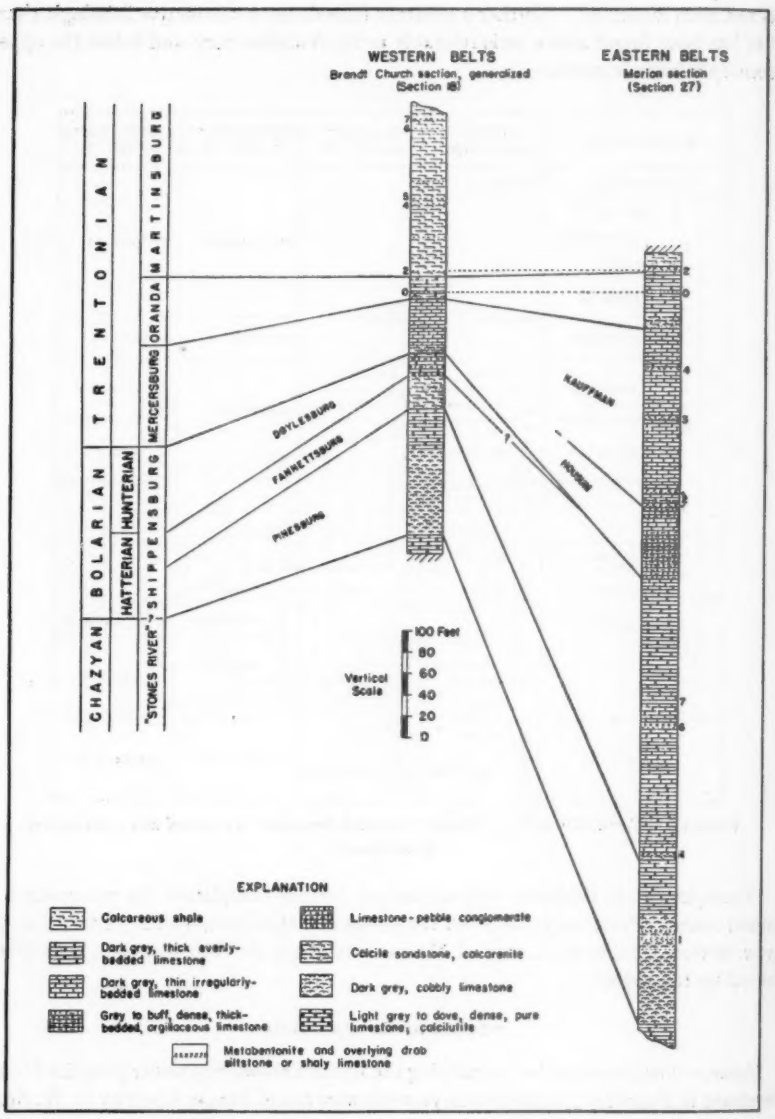


FIGURE 2.—*Relation of Bolarian and Trentonian formations between western and eastern belts of outcrop, south-central Pennsylvania*

Shippensburg formation. In the absence of another name, Stones River will be used here in the original sense, but in quotation.

As previously indicated, 150 feet of *Tetradium cellulosum* beds were referred from the "Stones River" to the Chambersburg of the type section (Stose, 1909, p. 8). Later, Butts, Stose, and Jonas (1932, p. 16) referred to the beds as a separate formation, the Lowville limestone. Cooper and Cooper (1946, p. 55) have shown that these beds are lithologically indistinguishable from the underlying 200 feet of the "Stones River" and identification as Lowville of New York seems unjustified. The "Stones River" limestone as used here refers to beds of the Beekmantown-Shippensburg interval and includes all of the *Tetradium cellulosum* beds at the Marion section.

Description.—A threefold division of the "Stones River" has been described (Cooper and Cooper, 1946, p. 69) at the type section of the Chambersburg: a lower dove calcilutite division with *Tetradium syringoporoides* Ulrich (30′ +); a middle, dark-grey, cherty limestone division with *Maclurites* (600′ +); and an upper, dark-grey to dove, fine-grained limestone division with *Tetradium cellulosum* (Hall) (350′ +). The detail of faunal and stratigraphic relations of the "Stones River" in south-central Pennsylvania and Maryland is beyond the scope of this study. The Shippensburg formation is consistently underlain by interbedded, dark-grey, fine-grained limestone and dove calcilutite in slabby to massive, blue- to white-weathering beds. *Tetradium cellulosum* (Hall) has been confidently identified in these beds only in the eastern belts of outcrop south of Chambersburg and in the Wilson belt in Maryland. Elsewhere a few questionable specimens of *Tetradium cellulosum* (Hall) have been found. However, *Tetradium syringoporoides* Ulrich was the most prominent fossil.

Correlation.—Beds like those of the Loysburg formation of central Pennsylvania have been recognized at Blacklog and Path Valleys and at St. Thomas (Kay, 1944, p. 6). A "tiger stripe" character, like that of the lower Loysburg formation, is present in the "Stones River" of the eastern belts in south-central Pennsylvania. Similarity of stratigraphic position and lithologic character indicate that the "Stones River" is at least partly equivalent to the Loysburg.

SHIPPENSBURG FORMATION

Definition.—The Shippensburg was originally defined (Craig, 1941b) as 500 feet of limestone overlying the "Lowville" of Butts, Stose, and Jonas (1932, p. 16). The Shippensburg formation is named for exposure at Middle Spring (section 26), 2.5 miles N.31°W. of Shippensburg, Franklin County. The formation is redefined as 416 feet of dark-grey, fine-grained limestone, cobbly below and thin crinkly to slabby above, underlain by the "Stones River" limestone and overlain by the Mercersburg formation. The type section (section 27) is along the Cumberland Valley Railroad, 2.1 miles S.53°W. of Marion, Franklin County.

The lower contact, a regional unconformity, is poorly exposed and marked only by contrasting lithologic character at the type section. The upper contact, representing a regional unconformity, is marked by lithologic change and a prominent, irregular, partially silicified bedding surface. In the western belts of outcrop, the formation is divided into three new members, Pinesburg, Fannettsburg, and Doylesburg. In the eastern belts, the Shippensburg attains its greatest thickness, yet only the lower two members, Pinesburg and Fannettsburg, have been recognized. In order that stratigraphic relations may be clear and independent of correlations, all

three members are defined in a single section near St. Thomas in the western belts. Because of the absence of satisfactory names at the type section, the members are named for more distant localities near which they are well displayed.

Lithologic characteristics.—The Shippensburg contains five distinct limestone types in south-central Pennsylvania and adjacent Maryland; dark grey, fine grained and cobbly; dark grey, fine grained and evenly bedded; medium to coarse calcarenite; interbedded dark-grey, medium-grained limestone and calcilutite pellet conglomerate; and dove calcilutite. These will be considered in discussing the members.

Seven metabentonites have been recognized in the Shippensburg of the eastern belts of outcrop and have been numbered from bottom to top. All the metabentonites have not been identified in any single section, and the sequence has been established by correlating three key beds which have either distinctive associated lithologic characteristics, distinctive faunal relations, or both. Intermediate metabentonites were correlated largely on stratigraphic position and the correlations are regarded as tenuous.

Thickness.—The 416 feet of Shippensburg at the type section thins to the northeast, south, and west. Northeast, thicknesses range irregularly from 181 feet at Newville (section 24) to 225 feet at Middle Spring (section 26). South, the Shippensburg is 266 feet thick at Wilson, Maryland (section 20). West, it thins abruptly to 165 feet west of St. Thomas (section 15), is 82–95 feet at Fort Loudon (section 10), and reaches minima of 60–65 feet and 72 feet at Willow Hill (section 7) and Spring Run (section 6) on the eastern limb of the Path Valley anticline. Thickening to the west, equivalent beds are 98 feet in Blacklog Valley.

Isopachs (Pl. 3) trend north from Wilson, Maryland, to the latitude of Marion where they bear northeast. The isopachs trend east of the northeast strike except between Shippensburg and Plainfield where they may roughly paralel the strike. Between Marion and Williamson, the isopachs are particularly crowded. Thickness variations reflect a combination of westward thinning by convergence and overlap and an absence of beds to the east by unconformity.

The fauna and the correlation of the Shippensburg will be discussed by members.

PINESBURG MEMBER OF THE SHIPPENSBURG FORMATION

Definition.—The Pinesburg member (new) of the Shippensburg formation consists of 114 feet of dark-grey, fine- to medium-grained limestone, cobbly below and platy above, with the type section on the south side of U. S. Route 30, 1.0 mile S.70°W. of St. Thomas, Pennsylvania (section 15). It lies disconformably on the dark-grey to dove, *Tetradium*-bearing "Stones River" limestone, and is overlain conformably by the Fannettsburg member of the Shippensburg. The member is named for exposure along the Western Maryland Railroad at Pinesburg Station, Maryland.

Lithologic characteristics.—At the type section, the lower 15 feet are dark-grey to black, medium-grained, dark platy weathering limestone with irregular partings. The succeeding 63 feet consists largely of dark-grey, fine-grained, cobbly, blue-weathering limestone with prominent thick silty partings, and contains numerous specimens of *Nidulites pyriformis* Bassler, *Dinorthis transversa* Willard, and *Sowerbyella* sp. cf. *S. aequistriata* (Willard). The upper 36 feet of the member is dark-grey, fine- to medium-grained, platy to slabby limestone.

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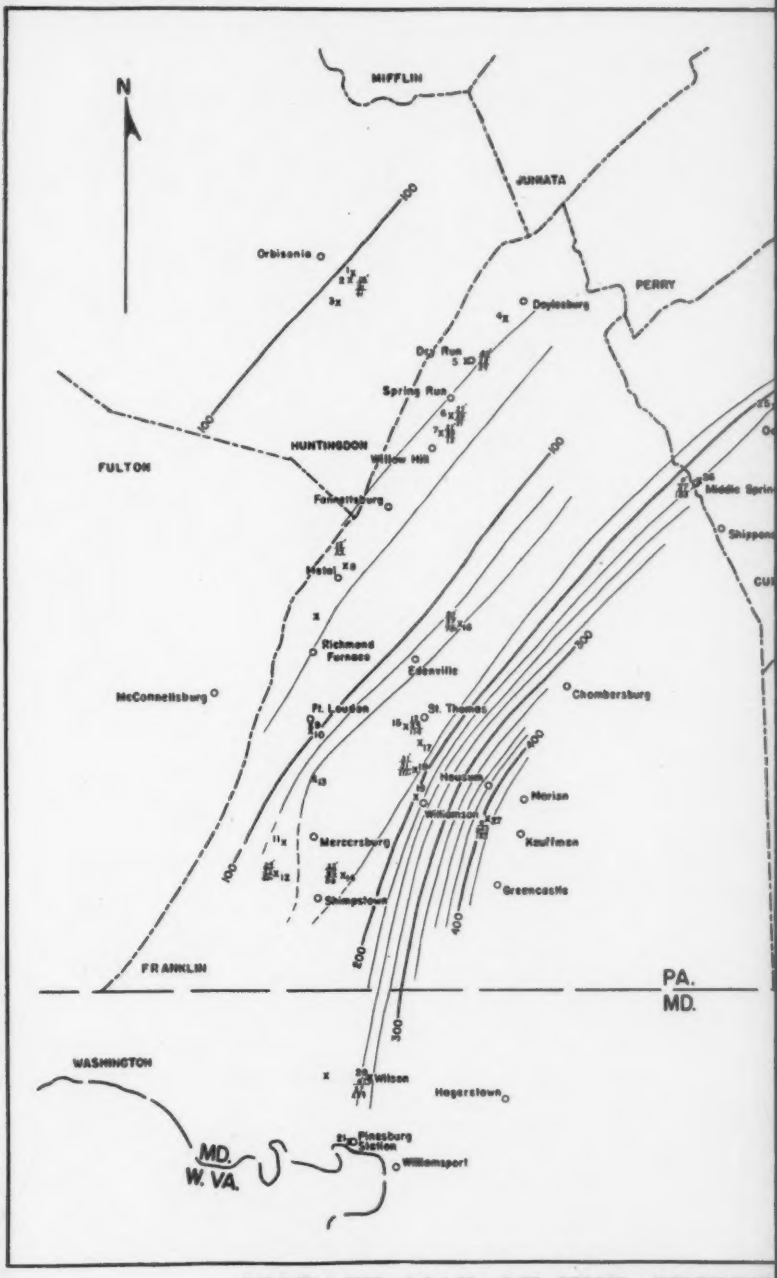

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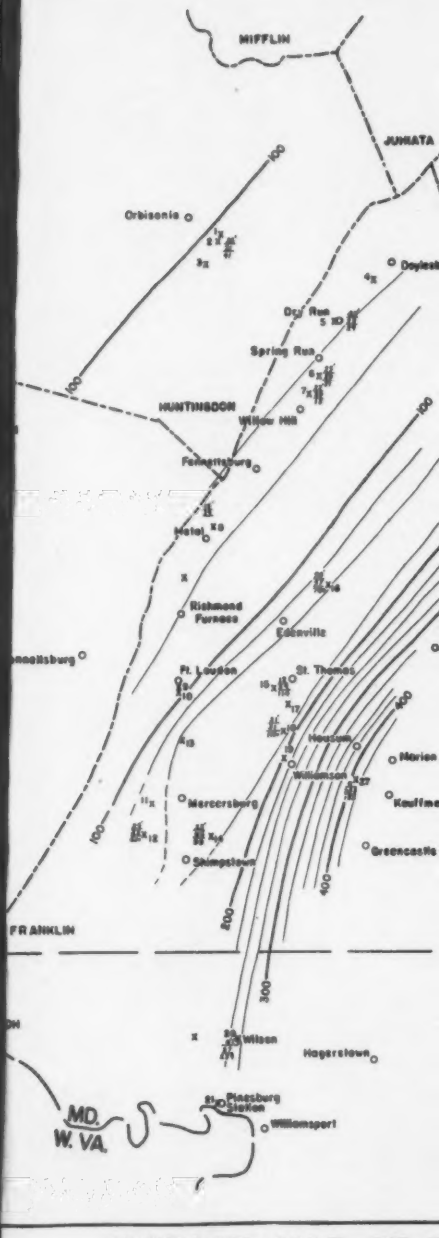

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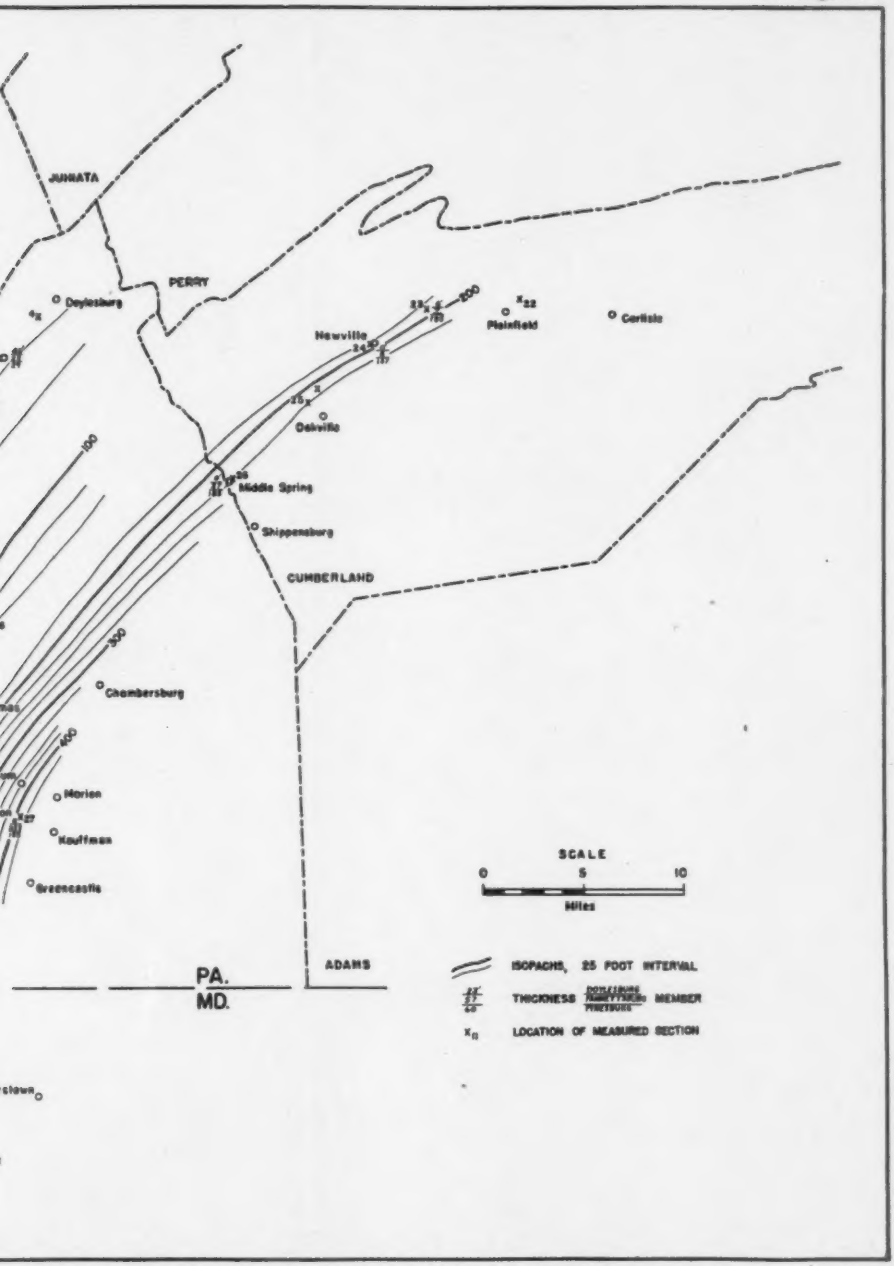

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A similar lithologic zoning has been recognized throughout the Cumberland Valley. The coarser-textured lowest part is the thinnest and least distinct unit and may represent a basal transgressional facies. At Dickey (section 13) it is represented by 3 feet of basal calcarenite beneath the middle cobbly zone. A similar basal calcarenite is present in Path Valley beneath the platy upper zone near Spring Run (section 6). Elsewhere the lower zone is indistinct and grades into the overlying cobbly beds. The cobbly beds form most of the lower part of the member in the Cumberland Valley but are absent by convergence and overlap in northern Path Valley. At Kough Quarry near Newville (section 24) 20 feet of medium to coarse calcarenite occurs within the cobbly beds. It has not been recognized elsewhere and is interpreted as a lens of local extent. The more evenly-bedded upper part of the member is present throughout the area. In Blacklog and Path Valleys and at Fort Loudon it contains abundant *Girvanella*. In many of the western belt sections the platy character is not prominent but in the eastern belt sections the zone is distinct and in places exhibits quite massive beds.

The Pinesburg member contains Shippensburg metabentonites 1, 2, and 3. Metabentonite 1 refers to a group of metabentonitic beds, present at the type section of the Pinesburg as thin clay partings 53 and 63 feet above the base of the member. The lower of these is a single clay bed while the upper consists of two clay beds separated by 7 inches of limestone and shale. At Pinesburg Station, Maryland, (section 21) the group consists of seven clay or shale partings within 39 feet of beds; at Wilson, Maryland, (section 20) six partings within 20 feet; and at Marion (section 27) three partings within 8 feet. Except for a possible correlation between the top beds at Marion and at St. Thomas, both double in character, none of the individual beds can be correlated between sections. Correlation as a group is warranted by the similar character of the beds themselves, their association with the cobbly beds and with the abundant occurrence of *Dinorthis transversa* Willard and *Echinosphaerites*. Shippensburg metabentonite 1 is present in all eastern belt sections from Middle Spring southward. In western belts it has been observed only at the section 1 mile west of St. Thomas. However, the poor exposure that characterizes the lower part of the Pinesburg in the western belts may obscure the true distribution.

Metabentonites 2 and 3 have been observed at only three sections. At Wilson, Maryland (section 20), two metabentonites, present in the upper platy to slabby part of the Pinesburg, are designated Shippensburg 2 and 3. At Kough Quarry, Newville (section 24), two metabentonites are in the platy upper Pinesburg above the abundant occurrence of *Dinorthis transversa* Willard and below Shippensburg metabentonite 4 which marks the base of the Fannettsburg member. In the absence of other field evidence, these are referred to as Shippensburg 2 and 3 on the basis of stratigraphic position. At the poorly exposed and structurally complicated section 1 mile northwest of Oakville (section 25), Shippensburg 2 and 3 have been tentatively identified by comparison with the Kough Quarry section.

Thickness.—The Pinesburg member is 153 feet thick at Marion, and thickens both to the south and the northeast in the eastern belts of outcrop. At Willson, Maryland, it is 238 feet thick, and between Shippensburg and Plainfield varies irregularly from 173 feet to more than 203 feet. The member thins uniformly from the eastern belts of outcrop to Path Valley. The 153 feet at Marion thins to 114

TABLE 4.—*Faunules from the Pinesburg member, Shippensburg formation**

	Localities									
	1	2	3	4	5	6	7	8	9	10
Girvanella sp.	c	.	.	.	.	.	.	.	c	.
Nidulites pyriformis Bassler	.	.	c	c	.	c	c	c	c	c
Receptaculites sp.	.	.	.	u	c	.	u	.	.	.
Lambeophyllum sp. cf. *L. profundum* (Conrad)	.	.	.	u	.	u	.	.	.	u
Tetradium cellulosum (Hall)	.	.	.	r	.	.	.	.	.	.
Echinosphaerites aurantium (Gyllenhal)	.	.	.	.	.	a	c	.	.	.
Ceramoporella (?) sp.	.	.	.	.	.	.	.	.	r	.
Diplotrypa ? appalachia Bassler	.	.	.	.	.	c	.	c	.	c
Escharopora (?) sp.	.	.	.	.	.	.	.	.	u	.
Hemiphragma (?) sp.	.	.	.	.	.	.	.	.	r	.
Monotrypa sp.	.	.	.	.	.	u	u	.	.	.
Rhinidictya sp.	.	.	.	.	.	.	.	.	c	.
Subretopora (*Chasmatopora*) sp.	.	.	.	.	.	.	.	.	r	.
Camerella sp. cf. *C. varians* Billings	r	.	.	.	.	.	.	.	.	.
Dinorthis transversa Willard	.	.	.	c	c	c	a	c	u	.
Doleroides (?) sp.	.	.	.	.	.	c	.	.	c	.
Glyptorthis sp. cf. *G. bellarugosa* (Conrad)	.	.	.	.	u	u	.	.	.	.
Hesperorthis sp. cf. *H. tricenaria* (Conrad)	.	.	.	u	c	.	.	.	u	.
Leptaena sp.	.	.	.	.	r	.	.	.	.	.
Leptellina (?) sp. cf. *L. elegantula* Butts	.	.	.	u	.	r	.	.	.	.
Lingula spp.	.	.	.	.	u	u	.	r	u	.
Lingulasma sp.	.	.	.	.	.	u	r	r	.	.
Mimella sp.	.	.	.	u	.	.	u	.	c	.
Multicostella sp.	.	.	.	.	.	.	r	.	r	.
Öpikina sp. cf. *Ö. ruedemanni* Salmon	.	.	.	.	u	.	.	r	u	.
Öpikina sp. aff. *Ö. septata* Salmon	.	.	.	u	.	.	.	.	u	.
Öpikina sp. cf. *Ö. wagneri* (Okulitch)	u	c	u	r	u	.	.	u	c	.
Oxoplecia sp. aff. *O. calhouni* Wilson	.	.	.	.	.	.	r	.	.	.
Oxoplecia simulatrix (Bassler)	.	.	.	.	.	u	r	.	.	.
Parastrophina sp. cf. *P. rotundiformis* (Willard)	.	.	.	.	r	.	.	.	r	.
Protozyga sp.	.	.	.	.	.	.	.	.	u	.
Resserella sp.	.	.	.	.	.	u	u	.	u	.
Skenidium (?) sp. cf. *S. anthonense* (Sardeson)	.	.	.	.	.	r	.	.	.	.
Sowerbyella sp. cf. *S. aequistriata* (Willard)	.	c	.	c	.	u	c	a	a	.
Sowerbyella sp. cf. *S. alternata* Butts	.	.	.	c	.	.	r	.	u	.
"*Sowerbyella*" sp. cf. "*S.*" *pisum* (Ruedemann)	.	.	.	.	c	c	.	.	r	.
Strophomena sp.	.	c	c	c	c	.	u	u	r	.
Strophomena sp.	.	.	.	.	.	.	.	.	r	.
Strophomena sp.	u	.	.	.	.	.	.	.	.	.
Bucania (?) sp.	.	.	.	.	r	.	.	.	.	.
Clathrospira (?) sp.	.	.	.	r	.	.	.	.	.	.
Hormotoma sp.	.	.	.	r	.	.	.	.	.	.

TABLE 4.—*Continued*

	Localities									
	1	2	3	4	5	6	7	8	9	10
Lophospira sp. cf. *L. bicincta* (Hall)	.	.	.	r	.	.	.	.	.	.
Lophospira sp.	.	.	.	r	.	.	.	.	r	.
Salpingostoma sp.	.	.	.	.	.	r	.	.	.	.
Calliops sp. cf. *C. annulatus* (Raymond)	c	.	.	r	.	u	u	.	u	.
Calliops sp. aff. *C. callicephalus* (Hall)	.	.	.	.	.	.	r	.	u	.
Illaenus sp. cf. *I. americanus* Billings	.	.	.	.	.	r	.	.	.	.
Illaenus sp. cf. *I. fieldi* Raymond	.	.	.	.	.	.	.	.	u	.
Illaenus spp.	.	.	.	.	r	r	.	.	u	.
Isotelus sp.	.	.	.	c	.	a	c	.	u	.
Thaleops sp.	r	.	.	.	.	.	.	.	.	.

* a, abundant; c, common; u, uncommon; r, rare. Localities: 1, Fort Loudon field section, one-half mile south of Fort Loudon; 2, Blue Spring section, 2.8 miles S.54°W. of Mercersburg; 3, Shimpstown section, along West Branch of Conococheague Creek, 2.5 miles southeast of Mercersburg; 4, St. Thomas section, along U. S. Route 30, 1.0 miles S.70°W. of St. Thomas; 5, Fry Coal and Stone Company Quarry, Williamson; 6, U. S. Route 40 at Wilson, Maryland; 7, Marion section, along Cumberland Valley Railroad, 2.1 miles S.53°W. of Marion; 8, field and road cut exposures at Middle Spring; 9, Kough Quarry, western environs of Newville; 10, Pennsylvania Turnpike, 2.9 miles N.45°E. of Newville.

feet at the type section west of St. Thomas, 89 feet near Shimpstown (section 14), 70 feet northeast of Edenville (section 16) and 41–64 feet at Fort Loudon (section 10). A minimum of 21 feet was measured near Spring Run on the east limb of the Path Valley anticline. Overlap of the lower beds and convergence throughout the member produce the thinning. The Pinesburg thickens westward from Spring Run to 29 feet at Dry Run (section 5) and equivalent beds are 41 feet thick in Blacklog Valley (section 2).

Fossils.—The Pinesburg member is relatively fossiliferous (Table 4), particularly in the eastern belts where it contains more abundant remains and a larger fauna than any of the succeeding units below the Martinsburg shale. The member corresponds to the *Echinosphaerites* zone and the lowermost part of the *Nidulites* zone (Stose, 1909, p. 9). *Echinosphaerites* was observed in the Pinesburg only at Marion and to the south. Stose (1908, p. 713), however, has listed *Echinosphaerites* plates from beds thought to be Pinesburg in western belts. *Nidulites* is present in the member throughout the eastern belts of outcrop, but has been found only at the St. Thomas section in the western belts. The apparent limited distribution of these fossils has probably contributed to past difficulties in correlation between eastern and western belts. *Dinorthis transversa* Willard and *Sowerbyella* sp. cf. *S. aequistriata* (Willard) are widespread.

Echinosphaerites, *Dinorthis transversa* Willard, *Sowerbyella* sp. cf. *S. aequistriata* (Willard), and *Isotelus* sp. are common associates. Their range of abundance in the Pinesburg is limited to the lower cobbly-bedded part. *Nidulites pyriformis* Bassler, *Diplotrypa ? appalachia* Bassler, and *Lambeophyllum* sp. cf. *L. profundum* (Conrad), common associates, range throughout the member and are the most prominent forms in the less fossiliferous upper part. *Girvanella* sp. is particularly abundant in the uppermost beds of the Pinesburg in Path Valley and has provided a useful zone for local correlation.

Correlation.—The Pinesburg member is equivalent to the Grazier member, and lower and middle parts of the Hostler member of the Hatter formation (Kay, 1941b, p. 1969; 1944, p. 10–11) in Blacklog Valley. The Eyer and Grazier members of the Hatter seem to overlap southeastward from Union Furnace in central Pennsylvania and only an upper part of the Grazier is present in Blacklog and Path Valleys (Kay, 1944, p. 10). The upper beds of the Pinesburg overlap the lower beds northwestward from the eastern belts of the Cumberland Valley to Path Valley. Similarity of stratigraphic position and depositional history suggests that Eyer and lower Grazier equivalents are present in the lower Pinesburg. No disconformity has been found in the Pinesburg that might correspond to that recognized by Kay (1944, p. 10) between the Eyer and Grazier at Union Furnace.

By faunal and lithologic similarity the Pinesburg member is equivalent to the lower part of the Edinburg formation at Strasburg, Virginia, (Cooper and Cooper, 1946, p. 78–86 and p. 95). The metabentonites at the base of the Edinburg at the Tumbling Run section near Strasburg probably represent Shippensburg 1. B. N. Cooper and G. A. Cooper (1946, p. 86) correlate the Edinburg with the Ward Cove-Witten interval (Cooper and Prouty, 1943, p. 863–879) in southwest Virginia. They also discussed the equivalence of the Virginia Chambersburg to the "Whitesburg" and "Athens" of the Harrisonburg area, Virginia (1946, p. 62–69), and to the Ottosee of Butts (1940, p. 170–178) in southwestern Virginia. Kay (1947, p. 1199) correlates upper Ward Cove with the Eyer and considers (1946, p. 1210) Benbolt to be similar to the Grazier and Hostler members. Thus, the Pinesburg probably corresponds to a part of the Ward Cove-Peery-Benbolt sequence of southwestern Virginia.

FANNETTSBURG MEMBER OF THE SHIPPENSBURG FORMATION

Definition.—At the type section (section 15), 1.0 miles S.70°W. of St. Thomas, the Fannettsburg member (new) consists of 39 feet of light- to dark-grey, medium-grained to coarsely crystalline limestone in platy to massive beds. In the western belts of outcrop, it is conformable between the underlying Pinesburg and the overlying Doylesburg members. In the eastern belts it lies conformably on the Pinesburg but is thought to be overlain disconformably by the Mercersburg formation. The Fannettsburg member is named for exposure east of Pennsylvania Route 75, 2.6 miles southwest of Fannettsburg in Path Valley.

Lithologic characteristics.—The Fannettsburg is dark and fine-grained in eastern belts and lighter, medium to coarsely crystalline in western belts (Fig. 2). A crude zoning has been recognized in Path Valley and elements of it persist through most of the western belts. The lower part of the Fannettsburg is typically medium-grey, rather thick-bedded, coarse calcarenite with layers of *Solenopora* conglomerate. The upper part is dark-grey, cross-laminated, fine calcarenite in platy beds. A thin interval of dove-grey, fine calcarenite is at the top in several of the westernmost sections. These lithic zones are gradational and locally are so interbedded that they cannot be separated.

In the eastern belts of outcrop, the Fannettsburg is dark-grey, fine-grained,

crinkly-platy to slabby limestone similar to both the underlying and overlying beds. At the Marion and Wilson, Maryland, sections, a few inches of fine calcarenite forms the base of the member. Shippensburg metabentonites 4 through 7 are within the Fannettsburg member. Their distribution is limited to the eastern belts of outcrop. Metabentonites 4, 5, and 6 are present at Wilson, Maryland, at 1.5, 6.3, and 14.2 feet respectively above the bottom of the Fannettsburg. Clay 4 is 2 inches of yellow to white metabentonite and brown shale. Six inches of platy, buff-weathering, argillaceous limestone with thick shale partings directly succeed the metabentonite. The distinctive lithologic character of this limestone forms the basis of correlation of clay 4 northward to Marion, Middle Spring, and Newville. Shippensburg metabentonite 6 is 2 inches of yellow to white metabentonite and drab shale overlain by 1 foot of slabby, cream to buff papery-weathering limestone that grades upward to the typical blue-grey weathering Shippensburg limestone. This metabentonite has been correlated to Marion on the basis of the associated distinctive limestone but has not been identified elsewhere in the eastern belts. Shippensburg 5 has no distinctive association and is designated at Wilson on the basis of stratigraphic position alone. A bed of metabentonite tentatively referred to Shippensburg 5 forms the highest Fannettsburg metabentonite at Middle Spring. It is 6 feet above Shippensburg 4 and exhibits no distinctive features. The bed may actually be Shippensburg 5, 6, or 7, or it may be a bed not present elsewhere; the designation is based on the similar thickness between Shippensburg 4 and 5 at Wilson and does not affect other correlations. Shippensburg metabentonite 7 has been identified only at Marion and exhibits no distinctive features or associations.

The base of the Fannettsburg is not synchronous through the area. In some western belt sections the lower contact is not sharp. Fine-grained limestone similar to the underlying Pinesburg is interbedded with Fannettsburg calcarenite. The base of the Fannettsburg has been arbitrarily taken as the lowest calcarenite. At Marion and Wilson, Maryland, in the eastern belts, thin beds of calcarenite have served as the base of the member. In other eastern belt sections the base of Shippensburg metabentonite 4 is taken as the bottom of the Fannettsburg for it is the lowest identifiable Fannettsburg horizon.

Lithologic correlation of the Fannettsburg between western and eastern belts is difficult because of the contrasting facies and the similarity of the eastern belt Fannettsburg to the underlying Pinesburg. However, a gradational intertonguing of the two facies may be observed between St. Thomas and Williamson (Pl. 2; Fig. 2). At the St. Thomas section (section 15), not quite half of the member is composed of coarse calcarenite, the rest is fine calcarenite. At the Brandt Church section (section 18), 2.7 miles S.17°W. of St. Thomas, less than one-third is coarse, and the remainder is medium-grained to finely crystalline laminated limestone. At Rockdale (section 19), half a mile northwest of Williamson, less than a quarter of the member is calcarenite, the remainder being dark-grey, fine- to medium-grained limestone like the eastern belt facies. This sequence of sections crosses belts of outcrop from west to east, and the Rockdale section is the easternmost section in the western belts of Pennsylvania. The thin, fine calcarenites at the base of the Fannettsburg at Marion and Wilson may represent the southeastern limit of the intertonguing.

TABLE 5.—*Faunules from the Fannettsburg member, Shippensburg formation**

	Localities									
	1	2	3	4	5	6	7	8	9	10
Solenopora sp.	a	.	a	a	.	.	.	u	.	.
Nidulites pyriformis Bassler	.	.	.	.	.	.	.	.	.	c
Cryptophragmus antiquatus Raymond	.	.	c	.	.	.	.	.	.	.
Stromatocerium sp.	c	.	.	.	.	.	.	.	.	.
Favistella halli	.	.	u	.	.	.	.	.	.	.
Caryocystitis (?) sp.	.		u	.	u	.	.	.	.	r
Carabocrinus (?) sp.	.	.	c	.	.	.	.	.	.	a
Diplotrypa ? appalachia Bassler	.	.	.	.	.	.	.	.	.	u
Escharopora sp.	c	.	c	.	c	.	.	.	.	u
Rhinidictya sp.	.	.	c	.	c	.	.	.	.	.
Subretopora (*Chasmatopora*) sp.	.	.	.	.	u	u	.	u	.	.
Campylorthis sp. cf. *C. magna* Schuchert and Cooper	u	.	.	.	.	.	.	.	.	r
Corineorthis sp.	.	.	.	.	.	.	.	.	.	r
Glyptorthis sp.	.	u	a	c	u	.	.	u	u	u
Hebertella sp.	.	.	.	.	.	.	.	.	.	r
Hesperorthis tricenaria (Conrad)	.	u	u	c	.	.	.	.	c	u
Leptaena ? sp.	.	.	c	.	c	c	c	a	c	.
Leptaena sp. cf. *L. homostriata* Butts	.	.	.	.	.	.	.	.	.	c
Leptobolus sp.	.	.	.	.	.	.	.	.	.	r
Mimella sp.	.	c	.	c	a	c	.	.	c	u
Multicostella sp.	.	r	r	.	.	.		.	.	.
Öpikina sp. aff. *Ö. ampla* Wilson	r	.	.	.	.	.	.	.	.	.
Öpikina sp. aff. *Ö. gloucesterensis* Wilson	c	.	.	.	.	.	.	.	.	.
Öpikina sp. cf. *Ö. wagneri* (Okulitch)	.	c	.	u	.	.	.	.	.	.
Oxoplecia simulatrix (Bassler)	.	.	.	.	.	.	.	.	.	u
Resserella sp.	.	.	.	.	.	.	.	.	.	u
cf. *Rostricellula plena* (Hall)	.	.	c	c	a	c	c	c	a	.
Sowerbyella sp. aff. *S. alternata* Butts	.	.	.	.	.	.	.	.	.	u
Sowerbyella sp. cf. *S. aequistriata* (Willard)	.	.	.	.	.	u	.	.	.	.
"*Sowerbyella*" sp. cf. "*S.*" *pisum* (Ruedemann)	.	.	.	.	.	.	.	.	.	r
Strophomena sp.	a	a	.	u	.	u	c	a	c	u
Zygospira recurvirostris (Hall)	.	.	c	c	a	a	.	.	a	.
Lophospira sp.	.	r	.	.	.	.	.	.	.	.
Calliops sp. cf. *C. annulatus* (Raymond)	.	.	.	.	.	.	c	.	.	c
Calliops sp. cf. *C. callicephalus* (Hall)	.	.	.	.	.	.	.	.	.	r
Homotelus sp.	.	.	.	.	.	u	.	.	.	c
Illaenus sp. cf. *I. americanus* Billings	.	.	.	.	.	.	.	.	.	u
Illaenus sp.	.	.	.	.	.	.	.	.	.	u
Lonchodomas sp. cf. *L. normalis* (Billings) and *L. halli* (Billings)	.	.	.	.	.	.	.	.	.	u

TABLE 5.—*Continued*

	Localities									
	1	2	3	4	5	6	7	8	9	10
Lonchodomas sp. cf. *L. sublaevis* Raymond	.	.	.	.	.	.	.	.	.	u
Tetralichas sp.	.	.	r	.	.	.	.	.	.	.
Thaleops sp.	u	.	c	a	c	.	.	u	c	.
Leperditia sp.	c	a	.	.	c	.	.	.	.	.
Macronotella sp. cf. *M. ulrichi* Ruedemann	.	.	.	a	.	.	.	.	.	.

* a, abundant; c, common; u, uncommon; r, rare. Localities: 1, Hammond Quarry, 0.8 miles S.5°E. of Spring Run; 2, Metal section east of Pennsylvania Route 75, 1.8 miles N.58°E. of Metal; 3, Fort Loudon Quarry, 0.4 miles south of Fort Loudon, west of Pennsylvania Route 75; 4, Africa section, 1.4 miles S.82°W. of Mercersburg; 5, Blue Spring section, 2.8 miles S.45°W. of Mercersburg; 6, Shimpstown section, along West Branch of Conococheague Creek, 2.5 miles southeast of Mercersburg; 7, Fort McCord section, 4.8 miles N.20°E. of St. Thomas; 8, St. Thomas section, along U. S. Route 30, 1.0 miles S.70°W. of St. Thomas; 9, Rockdale section, along Cumberland Valley Railroad, 0.5 miles N.45°W. of Williamson; 10, Marion section, along Cumberland Valley Railroad, 2.1 miles S.53°W. of Marion.

Thickness.—The Fannettsburg is 263 feet at the Marion section with Shippensburg metabentonites 4, 6, and 7 at 5, 124, and 148 feet respectively above the base of the member. At Wilson, Maryland, only 27 feet of beds is assigned to the Fannettsburg. Metabentonites 4 and 6 are 1½ and 14 feet from the base. Thus, between Marion and Wilson the beds between clays 4 and 6 converge from 119 feet to less than 13 feet. The upper beds of the Fannettsburg may show a similar convergence but this cannot be determined because of the disconformable relations interpreted at the top of the member.

To the northeast between Shippensburg and Plainfield, the Fannettsburg thins from 37 feet at Middle Spring (section 26) to 7 feet northeast of Newville (section 23). In this belt, clay 6 has not been identified. The Fannettsburg consists of beds between clay 4 and the disconformity at the base of the Mercersburg formation. Although the thickness of the Fannettsburg may reflect some convergence from Marion, it may also be a measure of beds removed during the post-Shippensburg hiatus.

To the west and northwest of Marion the Fannettsburg converges sharply and forms a thin irregular plate of beds. From the 263 feet at Marion, the member thins to 19 feet at Rockdale. It is 31 feet at Brandt Church and 38 feet west of St. Thomas. Western belt maxima of 57 and 55 feet are found northeast of Edenville (section 16) and at the Shimpstown section (section 14). At Dry Run in Path Valley (section 5) the Fannettsburg is 23 feet and in Blacklog Valley (section 2) equivalent beds are 21 feet thick.

Isopachs are approximately parallel to the present structural trend from Wilson, Maryland, to the latitude of Williamson, where they bear northeast across the structure. The thinning of the Fannettsburg northeast of Middle Spring deflects the isopachs northward across the northeast-trending belt of outcrop. The isopachs are especially crowded between Marion and Williamson where westward thinning is most pronounced.

Fossils.—Fossils (Table 5) are abundant in the western calcarenitic facies of the Fannettsburg but much less abundant in the eastern fine-grained limestones. None

of the common fossils appear to have a consistent and limited range of abundance within the Fannettsburg of the western belts. At the Marion section in the eastern belts of outcrop, *Caryocystites* (?) sp. and *Carabocrinus* (?) sp. are conspicuously associated and their range of abundance is restricted to about 10 feet in the middle of the member. The Fannettsburg of both eastern and western belts contains *Caryocystites* (?) sp., *Carabocrinus* (?) sp., and *Glyptorthis* sp. *Nidulites pyriformis* Bassler is common in the eastern belts but has not been observed in western belts. *Solenopora* sp. is characteristic of and abundant in the Fannettsburg of the western belts of outcrop. *Cryptophragmus antiquatus* Raymond has been found in the upper part of the Fannettsburg at Fort Loudon (section 9) and west of St. Thomas. The Fort Loudon occurrence was described by Ulrich (1911, p. 325–326) as immediately above a disconformity that formed the top of the *Caryocystites* zone and the bottom of the "Lowville," or *Tetradium cellulosum* zone. This stratigraphic break is well displayed at Fort Loudon but has not been recognized elsewhere in the western belts and is regarded as a local feature of diastemic magnitude. The limestones above and below the break are of similar character and have been included in the Fannettsburg. Thus, in the western belts, the member is equivalent to the upper part of the *Caryocystites* zone and the lower part of the *Tetradium cellulosum* zone of Ulrich. In the eastern belts the Fannettsburg is equivalent to a middle part of Stose's (1908, p. 712) and Ulrich's (Stose, 1909, p. 9) *Nidulites* zone.

Correlation.—By direct correlation in Blacklog Valley, the Fannettsburg is equivalent to the upper part of the Hostler member of the Hatter limestone and the lower part of the Snyder member of the Benner limestone (Kay, 1941b, p. 1969). The upper Hostler and lower Snyder are both granular, cross-laminated, evenly bedded; the contact arbitrarily taken at a prominent bedding plane is succeeded by a 3 inch bed containing white-weathering limestone pebbles. The horizon assumed as the contact between the Hostler and Snyder in Blacklog Valley satisfies the measurements presented by Kay (1944) and places the limestone pebble conglomerate, a characteristic of Snyder in central Pennsylvania, entirely within the Snyder. This contact does not mark the best separation of lithologic units in south central Pennsylvania, nor is it recognizable throughout the western belts of outcrop. The top of the Fannettsburg has been chosen at a higher horizon, the contact between granular limestone and overlying dove calcilutite. The member therefore contains some beds equivalent to lower Snyder of central Pennsylvania.

The lower granular Snyder is medium grey in the lower part, light grey to dove in the upper part, and is succeeded by dove calcilutite. This lithologic sequence has been observed in the upper Fannettsburg at Spring Run (section 6) and Metal (section 8) in Path Valley and at Fort Loudon to the south. Kay (1944, p. 17) has suggested that the Snyder overlaps southeastward from Blacklog Valley, but direct evidence has not been found; the lower Snyder equivalents are thin at Spring Run and do not necessarily form a synchronous unit.

The Fannettsburg is probably equivalent to an upper part of the Benbolt and a lower part of the Gratton-Bowen-Wardell-Witten interval (Cooper and Prouty, 1943) of southwestern Virginia. This is indicated by the correlation of the underlying Pinesburg with part of the Ward Cove-Peery-Benbolt interval and by the

identity of an upper part of the Fannettsburg with basal Snyder, which in turn has been correlated with the Gratton-Witten interval (Kay, 1947, p. 1199).

DOYLESBURG MEMBER OF THE SHIPPENSBURG FORMATION

Definition.—At the type section, 1.0 mile S.70°W. of St. Thomas (section 15) the Doylesburg member (new) is 12 feet of dove, massively-bedded, white-weathering calcilutite with 1 foot of dark-grey, fine-grained, platy, *Camarocladia*-bearing limestone at the top. *Cryptophragmus antiquatus* Raymond and *Tetradium cellulosum* (Hall) are the most abundant fossils. The Doylesburg is underlain conformably by the Fannettsburg member of the Shippensburg formation, and overlain disconformably by the Mercersburg formation. The member is named for exposure at the Binkley and Ober Quarry at Dry Run, 4 miles southwest of Doylesburg in Path Valley.

Lithologic characteristics.—In the western belts of outcrop the Doylesburg member exhibits two lithofacies (Pl. 2) which have a somewhat irregular distribution. The dove, thick-bedded calcilutite facies prevails from central Pennsylvania through Blacklog Valley and Path Valley. In the Cumberland Valley, this facies is found at the Shimpstown section (section 14), at a section 1 mile south of St. Thomas (section 17) and at the type section. The second facies is medium to dark-grey, fine- to medium-grained platy limestone with prominent heavy interbeds of medium-grey to dove limestone conglomerate. The conglomerate pebbles vary from tan and dove calcilutite to medium-grey, fine-grained limestone, and range from about one-sixteenth inch to 2 inches in maximum diameter. The great majority of the pebbles are less than one-half inch in diameter, and are rounded. The pebbles of larger dimensions have an irregular discoidal shape. This facies is best developed at the Brandt Church and Rockdale sections northwest of Williamson (sections 18 and 19). At Blue Spring (section 12), Fort Loudon (sections 9 and 10), and northeast of Edenville (section 16) tongues or lenses of the calcilutite facies occur within the darker, conglomerate-bearing facies. The facies variations are distributed in belts trending northeast across the more northerly trend of the present structure. These relations may be the fortuitous result of the few available samples but, if real, suggest depositional controls parallel to isopachs of the underlying units. The conglomerates are intraformational and were derived by disruption of slightly indurated limestone beds deposited in the area or in closely adjacent areas.

The contact of the Doylesburg with the underlying Fannettsburg is distinct in most cases. It is marked by sharp textural change from the granular Fannettsburg to the calcilutite, limestone conglomerate or dense, medium-grained limestone of the Doylesburg. The contact is either annealed or at a prominent bedding plane. The disconformity at the top of the Doylesburg shows slight relief, less than 6 inches, in several localities, and is marked typically by a succeeding basal calcarenite or basal semicrystalline limestone of the Mercersburg formation.

Thickness.—The Doylesburg varies irregularly in thickness through the western belts of outcrop, but shows an overall thinning from northwest to southeast. The maximum is 40 feet at the Binkley and Ober Quarry at Dry Run in Path Valley

(section 5). The minimum is 6 feet at the Rockdale section, northwest of Williamson. No clear correlation exists between the facies variations of the Doylesburg and the present thickness of the member. The irregularity is at least partly the result of post-Doylesburg erosion.

Determination of the distribution of the Doylesburg member in the eastern part of the area rests largely on interpretation. Absence of intermediate outcrop between the eastern and western belts as well as lithologic and faunal contrasts make direct field observations inconclusive. Three possibilities are suggested: (1) the Doylesburg is truncated to extinction southeastward and is absent in the eastern belts; (2) the member is present in the eastern belts by a gradational change in facies and fauna southeastward; (3) the member is present by reef separation of western lagoonal facies from an eastern benthonic facies.

A prominent, though quite regular, parting has been noted in several sections in the eastern belts within the interval between the highest known Shippensburg horizon and the lowest known Mercersburg horizon. In Kough quarry (section 24) this parting is overlain by a limestone bed containing a few dark chert nodules. At the Marion section the underlying surface is silicified. At Marion, Rocky Spring, and Kough quarry the parting marks a striking lithologic change and seems to represent the regional unconformity overlying the Doylesburg in the western belts of outcrop. The variation in thickness of interval between the known and correlatable horizons along the eastern belts, the variation in the actual beds delimiting this interval, and the apparent absence of both the *Caryocystites* horizons and Shippensburg metabentonites 5, 6, and 7 in the thinner sections are considered supporting evidence of the Shippensburg-Mercersburg unconformity in the eastern belts. The unconformity has not been definitely identified in several sections.

Two categories of evidence, direct and indirect, indicate the absence of the Doylesburg beneath this stratigraphic break in the eastern belts of outcrop. Direct evidence is negative in character. At Wilson the interval between the highest known bed of the Fannettsburg, Shippensburg metabentonite 6, and the lowest known bed of the Mercersburg, Mercersburg metabentonite 2, is 11.5 feet of dark-grey, slabby, relatively unfossiliferous limestone. These beds are lithologically identical to the beds beneath Shippensburg clay 6. Conclusion that the Doylesburg is absent at Wilson is based on the lack of any beds of typical Doylesburg facies, the apparent absence of characteristic Doylesburg genera, and the absence of any separate lithologic unit in this thin interval. The sections between Shippensburg and Plainfield present nearly the same relations. The interval between the highest known Fannettsburg bed, Shippensburg metabentonite 4 or 5, and the lowest known Mercersburg bed, Mercersburg metabentonite 1 is quite thin. No lithologic unit or characteristic faunal element of the Doylesburg is present. At Marion, 106 feet of dark-grey, slabby, relatively unfossiliferous limestone intervenes between the highest known Fannettsburg horizon, the highest observed *Caryocystites* occurrence, and the lowest known Mercersburg horizon, the base of a buff-weathering limestone member. These beds also lack any semblance of Doylesburg facies.

Indirect evidence is of positive character and is two-fold. The loss by truncation of more than 200 feet of beds between central Pennsylvania and Blacklog Valley,

and the thinning of the Doylesburg southeast from Blacklog Valley and Path Valley to a minimum of 6 feet at Rockdale, indicates consistently greater magnitude of disconformity eastward. Projection of this truncation to the eastern belts of outcrop in south central Pennsylvania suggests that the Doylesburg and possibly part of the Fannettsburg should be absent.

The light-colored calcilutite facies of the Doylesburg is a shallow-water deposit. The darker, dense, medium-grained limestone of the conglomerate-bearing facies appears to be a deeper-water and more normal marine limestone. The distribution of the two facies suggests that Doylesburg deposits originally thickened eastward from the western belts. Such thickening is not found in the proper interval in most eastern belt sections. The association of the dark, medium-grained marine limestone with limestone-pellet conglomerate may be explained by wave and current action in the westwardly shallowing water.

Argument counter to this evidence is found in the similarity of the dense medium-grained limestone of the conglomerate-bearing facies of the Doylesburg to much of the Shippensburg and Mercersburg of the eastern belts. If this were its only lithofacies in the eastern belts, the Doylesburg might not be recognized as a unit distinct from the Fannettsburg, and the change in its character would probably be one of lithologic gradation from the western shallow-water calcilutite to an eastern, deeper-water, dark-grey limestone. A separation of the two lithofacies by a reef seems impossible, for both the Fannettsburg and the Doylesburg change to the dark-grey facies by lateral gradation in the easternmost of the western belts of outcrop.

The writer believes the balance of the evidence points to the absence of the Doylesburg in the eastern belts, but the possibility of thin equivalents cannot be catagorically denied.

Fossils.—The fauna of the Doylesburg (Table 6), particularly that of the calcilutite facies, is small. The most abundant forms are *Tetradium cellulosum* (Hall) and *Cryptophragmus antiquatus* Raymond, which are found in both facies. In the western belts the Doylesburg is approximately equivalent to Ulrich's *Beatricea* and *Tetradium cellulosum* zone (Stose, 1909, p. 9). The Doylesburg is more extensive than his *Beatricea* beds at Blue Spring and corresponds to only the upper part of his *Beatricea* and *Tetradium cellulosum* zone at Fort Loudon (Ulrich, 1911, p. 325–327). As previously indicated, the characteristic forms of the Doylesburg fauna are not represented in the Shippensburg of the eastern belts of outcrop. The *Tetradium cellulosum*-bearing beds of the upper part of the "Stones River" limestone at Marion are not equivalent to the Doylesburg of the western belts for the two units are separated by the Pinesburg and Fannettsburg members of the Shippensburg.

Correlation.—By direct comparison in Blacklog Valley, the calcilutite beds forming the upper part of the Snyder member of the Benner limestone are equivalent to the Doylesburg member.

Cooper and Cooper (1946, p. 57–58) imply a correlation of the *Cryptophragmus*-bearing St. Luke member of the Edinburg with *Cryptophragmus*-bearing beds in the western belts of south-central Pennsylvania. If this correlation is correct and the St. Luke is the uppermost unit of the Edinburg, the Mercersburg formation or equivalent beds must be absent in northwestern Virginia, for the St. Luke member

is directly overlain by the Oranda formation. However, Cooper and Cooper (1946, p. 86) state the possibility that the Edinburg contains early Trentonian equivalents. If the St. Luke is the uppermost unit of the Edinburg then it should be early Tren-

TABLE 6.—*Faunules from the Doylesburg member, Shippensburg formation**

	Localities										
	1	2	3	4	5	6	7	8	9	10	11
Camarocladia sp.	.	.	.	.	.	.	.	.	u	.	.
Cryptophragmus antiquatus Raymond	.	c	.	.	a	a	c	a	c	a	.
Prismostylus columnaris (Hall)	c	.	.	.	.	.	.	.	.	.	.
Tetradium cellulosum (Hall)	c	c	.	a	a	c	a	.	c	.	.
Tetradium fibratum Safford	a	c	a	.	.	.	.	.	.	.	.
Tetradium syringoporoides Ulrich	.	.	.	.	.	.	.	c	.	.	.
Mimella sp.	.	.	.	.	.	.	.	.	.	.	u
Zygospira recurvirostris (Hall)	.	.	.	.	.	a	.	.	.	.	.
Helicotoma sp. cf. *H. planulatoides* Ulrich	.	.	.	.	c	.	.	.	.	.	.
Helicotoma sp. cf. *H. verticalis* Ulrich	.	.	.	.	a	.	.	.	.	.	.
Bathyurus sp.	.	.	.	.	.	.	.	r	.	.	.
Homotelus sp.	.	.	.	.	.	.	.	.	.	r	.
Illaenus sp.	.	.	.	.	.	.	.	.	.	r	.
Thaleops sp.	.	.	.	.	.	.	.	.	.	r	.
Leperditia sp.	.	.	.	.	.	.	.	c	.	a	.

* a, abundant; c, common; u, uncommon; r, rare. Localities: 1, Snyder member, Benner formation, Grove (Binkley and Ober) Quarry section, 1.8 miles S.54°E. of Orbisonia, Pa.; 2, Dry Run (Binkley and Ober) Quarry section, southeast environs of Dry Run; 3, Metal section east of Pennsylvania Route 75, 1.8 miles N.58° E. of Metal; 4, Fort Loudon Quarry, 0.4 miles south of Fort Loudon, west of Pennsylvania Route 75; 5, Fort Loudon field section, one-half mile south of Fort Loudon; 6, Blue Spring section, 2.8 miles S.54°W. of Mercersburg; 7, Dickey section, Cumberland Valley Railroad, 0.6 miles N.30°W. of Dickey; 8, Fort McCord section, 4.8 miles N.20°E. of St. Thomas; 9, St. Thomas section, along U. S. Route 30, 1.0 miles S.70°W. of St. Thomas; 10, Brandt Church section, 2.7 miles S.13°W. of St. Thomas; 11, Rockdale section, along Cumberland Valley Railroad, 0.5 miles N.54°W. of Williamson.

tonian and cannot correlate with the *Cryptophragmus*-bearing beds of south-central Pennsylvania which are here considered to be pre-Trentonian.

Cyptophragmus antiquatus Raymond (*Beatricea gracilis*) has been used as a guide fosssil in Middle Ordovician correlations throughout the eastern United States and part of Canada. However, it ranges through an appreciable thickness of beds. Cooper and Cooper (1946, p. 58) report it as low as 35 feet above the *Echinosphaerites* beds of the Edinburg formation near Wadesville, Virginia. Okulitch (1939, p. 336) reports it as high as the Coboconk limestone in Ontario, which has been variously interpreted as Chaumont and Rockland in age (Johnston, 1911, p. 189; Kay, 1937, p. 256–257; Kay, 1942a, p. 599–604; Okulitch, 1939, p. 336). In terms of the south-central Pennsylvania sequence, this range is approximately from middle Pinesburg to, at most, lower Mercersburg. However, the abundant occurrence of

Cryptophragmus antiquatus Raymond seems fairly restricted. In south-central Pennsylvania, the abundant occurrence is in the Doylesburg; in central Pennsylvania it is in the Stover member of the Benner formation (Kay, 1944, p. 22). In southwestern Virginia the abundant occurrence is in the Witten limestone (Cooper and Prouty, 1943, p. 877–879). In northwestern Virginia a relatively abundant occurrence is in the St. Luke member of the Edinburg formation. Cooper and Cooper (1946, p. 59) have tentatively correlated the St. Luke with the Witten. Kay (1947, p. 1199) indicates that the Witten contains at least partial equivalents of Snyder and Stover as well as Nealmont.

MERCERSBURG FORMATION

Definition.—The Mercersburg formation is named for exposure along the West Branch of Conococheague Creek, 2.5 miles S.38°E. of Mercersburg (section 14), but the type section is along the Cumberland Valley Railroad and in adjacent fields, two miles S.53°W. of Marion (section 27). The Mercersburg was defined (Craig, 1941b, p. 1964) as 150 feet of dark-grey, thin irregular- and thick even-bedded limestone with the base at the bottom of a drab cuneiform siltstone zone containing a metabentonite. Later field work has indicated the need for redefinition. The major change is the addition to the Mercersburg of a unit of beds from the top of the original Shippensburg. The Mercersburg formation is redefined as 235 feet of dark-grey, fine-grained, thin crinkly- and thick evenly-bedded blue-grey weathering limestone with an interval of tan to medium-grey, slabby, buff-weathering limestone at the base. It is underlain by the Shippensburg formation and overlain by the Oranda formation.

The lower contact, representing a regional unconformity, is marked at the type section by a prominent, irregular, partially silicified surface separating the buff-weathering lower Mercersburg from the blue-grey weathering upper Shippensburg. The upper contact, a minor disconformity, is marked by the prominent basal calcarenite of the Oranda formation. In several sections the uppermost bedding plane of the Mercersburg exhibits slight pitting.

Lithologic characteristics.—Along the eastern belts of outcrop in south-central Pennsylvania, distinctive lithologic differences form the basis for separating the Mercersburg formation into two members: a lower, grey to tan, fine-grained, argillaceous, slabby, buff-weathering limestone and an upper, dark-grey, fine- to medium-grained, thin crinkly to slabby, blue-grey weathering limestone. Some interbedding of the two rock types has been observed locally. In the western belts of outcrop, the Mercersburg is dark-grey, fine- to medium-grained, thin crinkly limestone quite similar to the upper member of the eastern belts and in most places exhibits a basal calcarenite with a maximum observed thickness of 8 feet.

Five metabentonite beds are present in the Mercersburg and have been correlated with varying degrees of certainty. Nowhere have all of them been found in one section. The clays are numbered from one to four in ascending order. Mercersburg 2 is a double metabentonite in several sections.

Thickness.—The Mercersburg formation thins uniformly to the northeast from the 235 feet at Marion (Pl. 4). It is 155 feet at Middle Spring (section 26), 130 feet

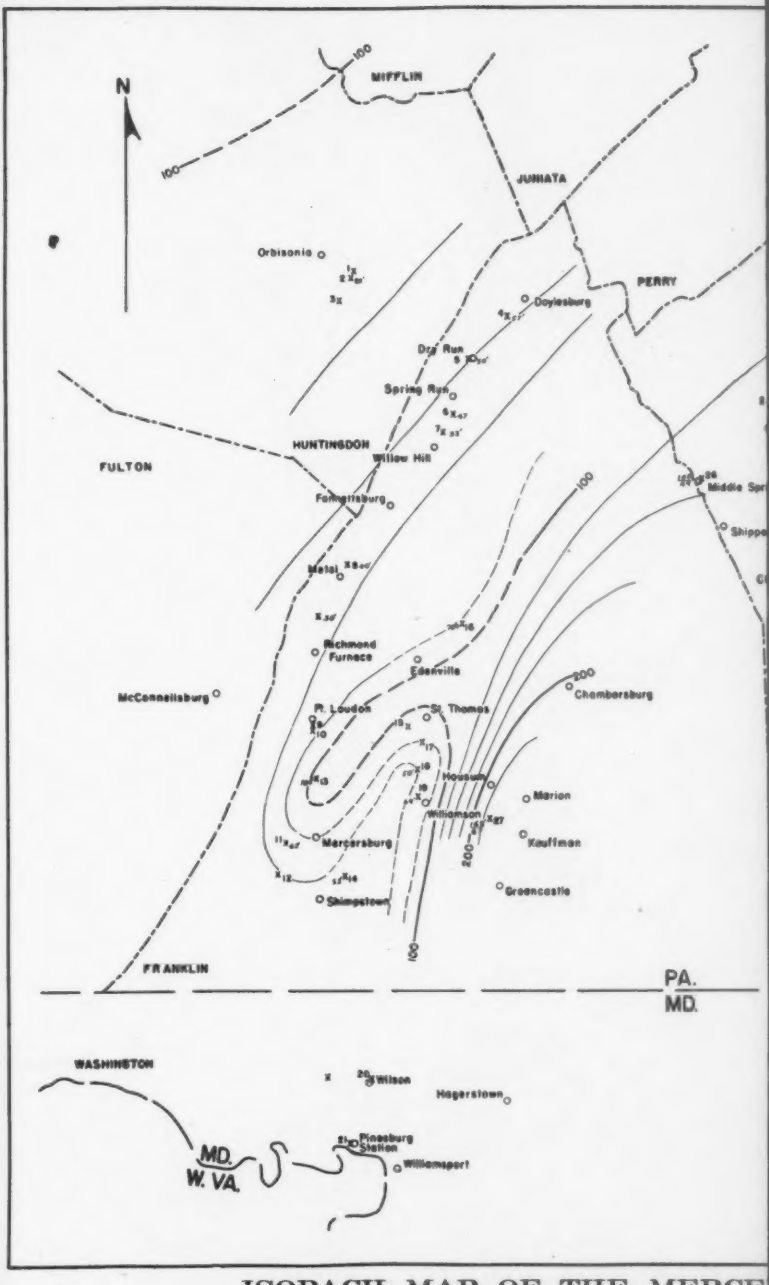
northeast of Newville (section 23) and somewhat more than 108 feet northeast of Plainfield (section 22). It thins abruptly from Marion to 69 feet and 50 feet northwest of Williamson (sections 18 and 19). From Williamson, it thins irregularly west and northwest, reaching a minimum thickness of 33 feet at Willow Hill (section 7) in Path Valley and thickening again to 84 feet of equivalent beds in Blacklog Valley (section 2).

The fauna and correlation of the Mercersburg formation will be considered by members.

HOUSUM MEMBER OF THE MERCERSBURG FORMATION

Definition.—The type section of the Housum member (new) of the Mercersburg formation is along the Cumberland Valley Railroad 2 miles southwest of Marion (section 27). At this section the Housum is 67 feet of tan to medium-grey, fine-grained, argillaceous, slabby, cream- to buff-weathering limestone with a 10-foot interval of dark-grey, thin crinkly, grey-weathering limestone near the top of the member. The Housum is conformable beneath Mercersburg metabentonite 2 and is regarded as disconformable above the Fannettsburg member of the Shippensburg. The member is named for a small community 1.7 miles north of the type section.

Lithologic characteristics.—The distinctive buff-weathering Housum has been recognized only in the eastern belts of outcrop of south-central Pennsylvania. At the type section the 10-foot interval of grey-weathering limestone underlies 16 feet of buff-weathering limestone forming the top of the member. The bottom of the buff-weathering Housum is at a prominent, irregular, partially silicified bedding surface. At Rocky Spring Church, 3.7 miles N.15°W. of Chambersburg, the Housum is well exposed though the underlying beds are structurally complicated. Mercersburg metabentonite 2 is directly underlain by an interval of grey-weathering, crinkly-bedded limestone, which in turn is underlain by buff-weathering limestone containing Mercersburg metabentonite 1, 2 inches of yellow clay succeeded by 1 foot of olive-drab, fissile to blocky shale. Sharp lithologic change to dark-grey, thin, crinkly limestone marks the base of the Housum. From Middle Spring (section 26) northeastward to the Pennsylvania Turnpike near Newville (section 23) the upper part of this sequence is the same. However, the lithologic contrast marking the base of the Housum is less prominent. At Kough quarry (section 26) and at the Diller Church section (section 23) a prominent shaly parting a few feet below Mercersburg metabentonite 1 has been selected as the probable position of the Shippensburg-Mercersburg contact. At Marion and Rocky Spring, north of Chambersburg, a similar parting marks a striking lithologic change. No comparable parting was recognized at Middle Spring and near Oakville (section 25). Lacking other criteria, the base of the Housum has been arbitrarily placed at the base of Mercersburg metabentonite 1, which marks the most striking lithologic change in these sections. Confusion may result from too strict an interpretation of this arbitrary contact, assigned primarily to facilitate correlation. Mercersburg metabentonite 1 represents the lowest known bed of the formation. A small but unknown thickness of the underlying beds are Housum in age. The arbitrary contact should be interpreted as representing not the regional unconformity, for the preservation of a



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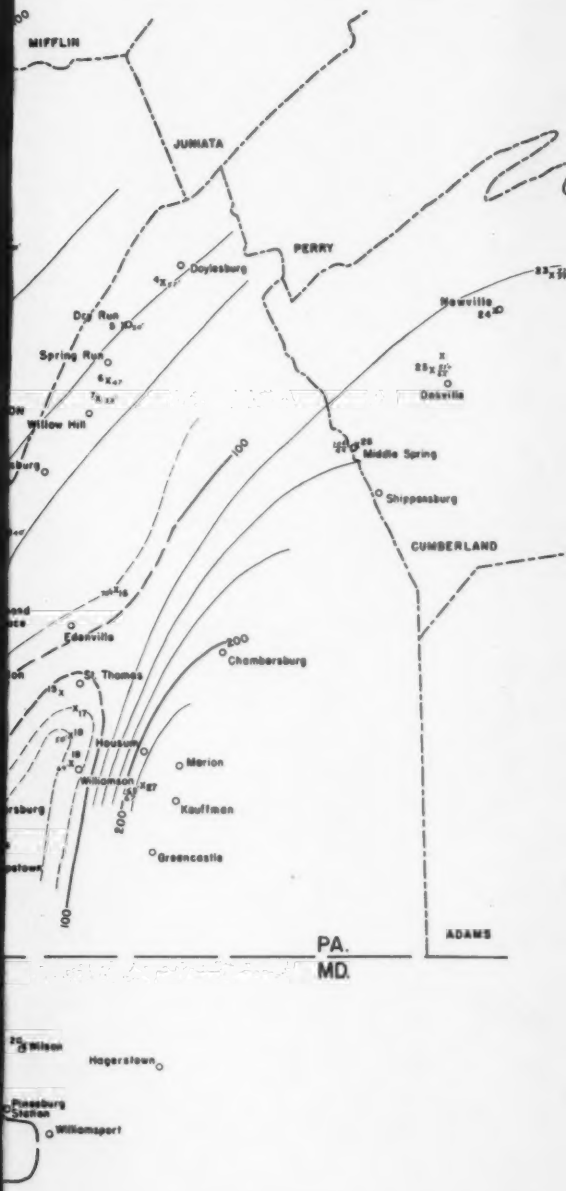
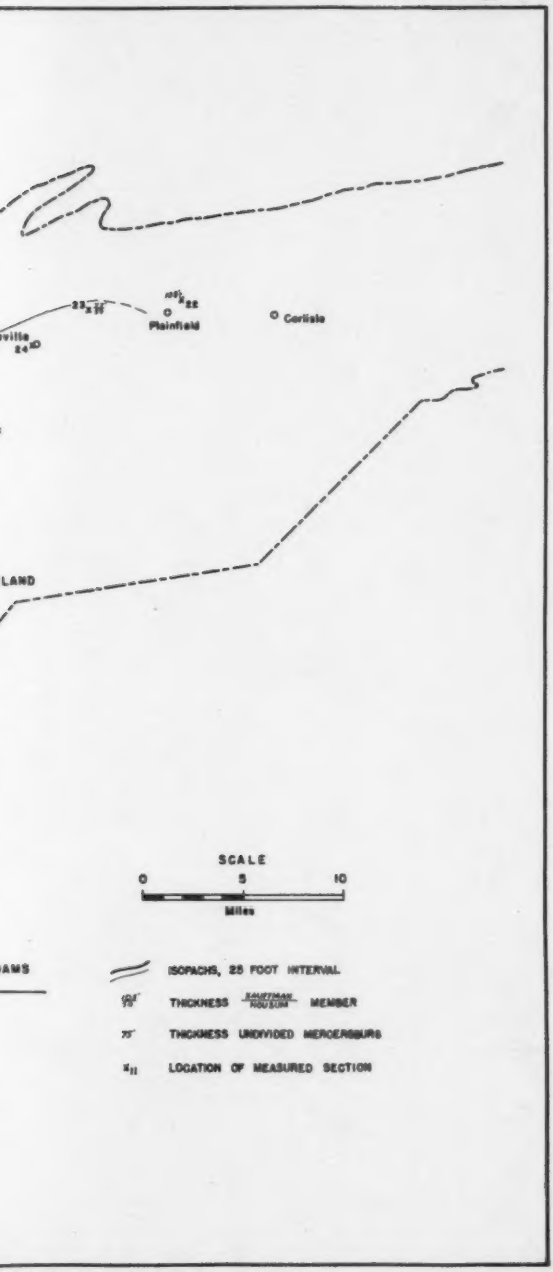
SOUTH-CENTRAL PEN



CH MAP OF THE MERCERSBURG
SOUTH-CENTRAL PENNSYLVAN



BURG FORMATION,
YLVANIA

metabentonite on such a surface would be virtually impossible, but the nearest location to the true contact possible under the field conditions. The prominent drab shale bed is present above metabentonite 1 as far as Oakville (section 25) but is absent in the vicinity of Newville (sections 23 and 24). Correlation of this metabentonite is based on its association with the buff-weathering limestone of the Housum. The drab shale bed is not a satisfactory criterion for correlation for it is not universally present and similar shales are associated with other Mercersburg metabentonites. Southward from Marion, the Housum has been tentatively identified along the Cumberland Valley Railroad in the northern environs of Greencastle. It is thought to be absent, or nearly so, at Wilson, Maryland, where the lowest identifiable horizon of the Mercersburg is clay 2.

The buff-weathering Housum is not present in the western belts of outcrop. Beds of Housum age are known only in sections containing metabentonite 2 and are dark-grey limestone with a variably thick basal calcarenite which rests disconformably on the Doylesburg member of the Shippensburg formation. Mercersburg metabentonite 1 has not been observed in the western belts of outcrop.

Thickness.—The Housum member thins northeastward from 67 feet at Marion to 29 feet at Rocky Spring. It is 49 feet at Middle Spring and 54 feet northeast of Newville.

In the western belts, beds equivalent to the Housum are 16 feet thick west of Mercersburg (section 11), 4 feet near Spring Run in Path Valley (section 6) and 26 feet in Blacklog Valley (section 2).

Fossils.—The Housum member contains the least fossiliferous beds of the "Stones River"-Martinsburg interval (Table 7). In the absence of faunal evidence, these beds were assigned originally to the Shippensburg (Craig, 1941, p. 1964). Subsequent discovery of *Leptaena* (?) sp. cf. *L. charlottae* Winchell and Schuchert, a brachiopod characteristic of the Mercersburg of the western belts and the Nealmont of central Pennsylvania, has indicated the proper disposition of the Housum member within the Mercersburg formation.

Although differences in measurement do not permit exact comparison with the early descriptions of the type section (Stose, 1908, p. 710; 1909, p. 8), the Housum is thought to correspond to an interval described as containing few fossils and having a position intermediate between the *Nidulites* and *Christiania* zones. In the western belts, the member corresponds stratigraphically to the third faunal zone or *Echinosphaerites* zone of Ulrich (*in* Stose, 1909, p. 9). The faunule listed for this zone is quite similar to the faunule listed earlier (Stose, 1908, p. 713) from a stratigraphically lower interval. No basis has been found for distinguishing an *Echinosphaerites* zone in the lower part of the Mercersburg formation in the western belts of outcrop.

Correlation.—The Housum member correlates directly with the lower part of the Nealmont formation (Kay, 1941b, p. 1969) in Blacklog Valley. If the single metabentonite in the Mercersburg formation in Blacklog Valley is Mercersburg 2 and if it corresponds to metabentonite N_2 at the top of the Oak Hall member of the Nealmont formation in central Pennsylvania, the Housum is equivalent to part or all of the Oak Hall. Synchroneity of the lower beds of the Housum and Oak Hall cannot be determined as both overlap on a regional unconformity.

KAUFFMAN MEMBER OF THE MERCERSBURG FORMATION

Definition.—The thickest and most complete section of the Kauffman member (new) is in the fields adjacent to the Cumberland Valley Railroad, 2.1 miles S.53°W. of Marion, Franklin County (section 27). Despite the lack of continuous exposure, this is selected as the type section. The Kauffman member is 168 feet of dark-grey, fine- to medium-grained, thin crinkly to slabby, grey-weathering limestone. It lies conformably above the Housum member of the Mercersburg formation and is disconformably below the Oranda formation. The member is named for the village of Kauffman, 1.6 miles S.65°E. of the type section.

Lithologic characteristics.—The Kauffman member forms the greater part of the Mercersburg formation throughout the area. In the western belts, it is usually platy-bedded with thin crinkly stylolitic partings. The beds weather to grey, roughened surfaces. In the eastern belts, this type of bedding and weathering predominates but intervals of lighter-grey, smooth-weathering limestone, either in thick slabby or thin cobbly beds, are locally developed.

Four metabentonites are present at 0, 15, 84, and 130 feet respectively in the Kauffman at the type section. Each clay bed is immediately succeeded by 1–2½ feet of olive-drab to green shale or blocky siltstone. The lower two metabentonite beds are referred to as Mercersburg metabentonite 2 because of their close association. The succeeding metabentonites are designated Mercersburg 3 and Mercersburg 4.

Correlation of the metabentonites in the Kauffman is difficult and several criteria must be used. In the eastern belts of outcrop Mercersburg metabentonite 2 is correlated by its position above the buff-weathering Housum, by its association with a thick, platy to blocky siltstone, and locally by its double character. West of Oakville (section 25), the lower clay bed of Mercersburg metabentonite 2 is succeeded by 8 feet of drab, blocky siltstone, 1 foot of grey limestone, and then the upper clay bed. At Middle Spring (section 26), the lower clay bed is succeeded by 8.5 feet of platy to blocky siltstone but the upper clay has not been found. Northeast of Newville on the Pennsylvania Turnpike (section 23) Mercersburg 2 is represented by a clay bed 28 feet above the buff-weathering facies in the Housum and is succeeded by one-half foot of siltstone and shale. Southward from Marion Mercersburg 2 has been tentatively identified in the railroad cut in the northern environs of Greencastle. Here a thin bed of cream-colored metabentonitic shale is underlain by buff-weathering limestone similar to that of the Housum and is overlain by about 20 feet of drab calcareous siltstone and shale. The upper clay was not discovered. At Wilson, Maryland, the Housum member has not been identified. The lowest bed assigned to the Mercersburg is a shaly metabentonite succeeded by 1½ feet of drab siltstone and shale. A second metabentonite bed is 14.5 feet higher. The similarity of this interval to that between the two clays of Mercersburg 2 at the type section and near Oakville is the tenuous basis for assigning the Wilson clays to Mercersburg 2. In this section, the base of clay 2 has been arbitrarily used as the base of the Mercersburg and 11.5 feet of beds immediately beneath the clay have been assigned to the Fannettsburg, as the exact horizon of the Shippensburg-Mercersburg disconformity in this subjacent interval cannot be located. Beds

underlying the arbitrary contact may be equivalent to Mercersburg. The contact should not be interpreted as representing the regional unconformity.

In the western belts of outcrop, Mercersburg metabentonite 2 is correlated by stratigraphic position and locally by associated siltstone beds. In Blacklog Valley (sections 1, 2, and 3) a single metabentonite is 26 feet above the base of the formation; at Dry Run (section 5) and near Spring Run (section 6) in northern Path Valley it is 6.5 feet and 4 feet respectively above the base. In the western part of the Cumberland Valley, only two sections show a metabentonite in the Mercersburg formation. At the Africa section (section 11), one mile S.82°W. of Mercersburg, the metabentonite is 15.5 feet above the base of the formation, and at the Dickey section (section 13) it is 6 feet above the base. At both Africa and Dickey, 2 feet of drab, shaly to blocky siltstone succeed the metabentonite. At Africa the metabentonite is double in character; the two clay beds are separated by one-half foot of drab shale. This double character as well as the blocky weathering of the associated siltstone form the bases for correlating this unit as Mercersburg metabentonite 2. In the eastern belts, Mercersburg metabentonite 2 is the only metabentonite unit to exhibit two closely spaced clay beds and blocky weathering in the associated siltstone.

Metabentonites 3 and 4 have been observed in the same section only at the type locality. Siltstone beds are associated with them only at the type locality, and correlation is based entirely on stratigraphic position. A metabentonite in the middle part of the Kauffman is referred to Mercersburg 3, and one near the top is referred to Mercersburg 4. Correlation by stratigraphic position is dangerous for the beds may be confused or an entirely different metabentonite may be miscorrelated, but it is used tentatively in the absence of other criteria. Thus, Mercersburg metabentonite 3 (?) is present at Wilson, Maryland (section 20), northwest of Williamson (section 19), and at the Pennsylvania Turnpike, northeast of Newville (section 23). Metabentonite 4(?) is present at Middle Spring.

Thickness.—The Kauffman member thins northeastward from 168 feet at Marion to 106 feet at Middle Spring, 107 feet on the Pennsylvania Turnpike, northeast of Newville, and 101 feet northeast of Plainfield (section 22). In the western belts the Kauffman may be distinguished only in those few sections containing Mercersburg metabentonite 2. The member thins to a minimum of 44 feet at Dry Run in Path Valley, thickening again to 59 feet in Blacklog Valley.

Fossils.—The Kauffman is relatively unfossiliferous southeast of northern Path Valley (Table 7). *Nidulites pyriformis* Bassler and *Diplotrypa ? appalachia* Bassler are the most common fossils in the eastern belts of outcrop. *Nidulites* is rare in the western belts, but has been tentatively identified from Willow Hill in Path Valley and has been reported (Kay, 1944, p. 106) from equivalent parts of the Nealmont formation at two localities in central Pennsylvania. The characteristic Mercersburg-Nealmont brachiopod, referred to as *Leptaena* (?) sp. cf. *L. charlottae* Winchell and Schuchert, is present in both eastern and western belts. Other more abundant brachiopods are *Sowerbyella* sp. cf. *S. punctistriata* Mather, *Hesperorthis tricenaria* (Conrad) and *Öpikina* spp.

In Path Valley two relatively abundant occurrences of fossils were noted in the

TABLE 7.—*Faunules from the Mercersburg formation and equivalent beds**

	Localities															
	1	2	3	4	5	6	7	8	9	10	11	12H	12K	13K	14H	14K
Girvanella sp.	.	.	.	.	.	.	c	.	.	.	.	.	.	u	.	.
Nidulites pyriformis Bassler	.	.	.	.	.	r	.	.	.	.	.	c	c	c	c	c
Lambeophyllum profundum (Conrad)	c	.	.	.	.	.	.	.	.	.	.	u	.	.	.	.
Streptelasma sp. cf. *S. corniculum* Hall	.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	r
Corynotrypa delicatula (James)	.	.	.	c	.	.	.	.	.	.	.	.	.	.	.	.
Diplotrypa ? appalachia Bassler	c	u	.	c	a	a	.	a	.	.	a	.	c	c	.	a
Escharopora sp.	.	.	.	.	c	.	.	.	.	.	.	.	.	.	.	.
Helopora sp.	.	.	.	.	c	c	.	.	.	.	.	.	.	.	.	.
Rhinidictya sp.	.	.	.	.	u	u	.	.	.	.	.	.	.	.	.	.
Stictoporella (?) sp.	.	.	.	.	.	.	.	u	u	.	.	.	.	.	.	.
Craniops sp. cf. *C. trentonensis* (Hall)	.	.	.	.	r	.	.	.	.	.	.	.	.	.	.	.
Dinorthis sp. cf. *D. pectinella* (Conrad)	.	r	.	.	c	c	.	.	.	.	.	.	.	.	.	r
Doleroides sp. cf. *D. pervetus* (Conrad)	.	u	c	.	r	.	.	.	.	.	.	.	.	.	.	r
Glyptorthis sp. cf. *G. bellarugosa* (Conrad)	.	.	.	.	.	.	.	r	.	r	.	.	r	.	.	.
Hesperorthis tricenaria (Conrad)	u	u	c	c	c	c	u	c	u	.	u	.	.	.	.	.
Leptaena sp. cf. *L. charlottae* Winchell and Schuchert	u	.	.	.	c	.	r	.	.	.	.	r	r	.	.	.
Leptaena sp. cf. *L. homostriata* Butts	.	.	.	.	r	.	.	.	.	.	.	.	.	.	.	.
Lingula sp.	.	.	.	.	.	.	.	.	.	.	.	.	.	.	r	.
Lingulasma sp.	.	.	.	.	.	.	.	.	.	.	.	u	.	.	.	.
Öpikina sp. cf. *Ö. minnesotensis* (Winchell)	.	.	a	a	.	.	u	.	c	u	u	.	.	.	.	.
Öpikina sp. aff. *Ö. ruedemanni* Salmon	.	.	.	.	.	.	.	r	.	.	r	.	.	.	r	.
Öpikina sp. aff. *Ö. subtriangularis* Wilson	.	.	.	u	a	a	.	.	.	.	.	.	.	.	.	.
Öpikina sp. cf. *Ö. transitionalis* (Okulitch)	.	.	.	r	.	.	.	.	.	.	.	.	.	.	.	.
Öpikina sp. cf. *Ö. wagneri* (Okulitch)	.	u	.	.	c	.	.	u	.	.	.	.	.	.	.	.
Ptychoglyptus sp.	.	.	.	.	.	.	r	.	.	.	.	.	.	.	.	.
Rostricellula sp. cf. *R. rostrata* Ulrich and Cooper	.	u	.	.	.	.	.	.	.	.	.	.	.	.	.	.
Skenidioides (?) sp.	.	.	.	.	r	.	.	.	.	.	.	.	.	.		.

TABLE 7.—*Continued*

	Localities															
	1	2	3	4	5	6	7	8	9	10	11	12H	12K	13K	14H	14K
Sowerbyella sp. cf. *S. alternata* Butts	.	r	r	.	.	r	.	.	.	.	.	.	.	.	.	.
"Sowerbyella" sp. cf. *"S." pisum* (Ruedemann)	.	.	.	.	c	c	.	.	.	.	.	.	r	.	.	c
Sowerbyella sp. cf. *S. punctistriata* (Mather)	.	a	a	c	.	.	c	.	.	.	c	.	c	c	r	.
Strophomena sp.	.	u	.	.	.	.	.	.	.	.	u	.	.	.	.	.
Strophomena sp.	.	.	.	.	u	.	.	.	u	.	.	.	.	.	.	r
Strophomena sp.	.	u	a	c	.	.	.	.	.	.	.	.	.	.	.	r
Triplesia sp.	.	.	.	.	.	.	.	c	c	.	.	.	.	r	.	.
Valcourea sp. cf. *V. loricula* (Hall)	.	.	.	.	c	u	.	.	.	.	.	.	.	.	.	.
Zygospira recurvirostris (Hall)	.	.	.	.	.	.	a	c	.	.	.	.	.	.	.	c
Ambonychia (?) sp.	.	.	.	.	.	.	.	.	.	r	.	.	.	.	.	.
Cyrtolites sp.	.	.	.	.	.	.	.	.	.	.	.	r	.	.	.	.
Phragmolites sp.	.	.	.	u	c	u	.	.	.	.	.	r	.	.	.	.
Tropidodiscus sp.	.	.	.	c	.	.	.	.	.	.	.	.	.	.	.	.
Bumastus sp.	.	.	u	.	.	.	.	c	.	u	.	u	.	.	.	.
Calliops sp. cf. *C. callicephalus* (Hall)	.	.	u	.	u	c	.	u	.	c	u	c	.	.	.	.
Ceraurus sp. cf. *C. dentatus* Raymond and Barton	.	.	.	.	u	.	.	.	.	.	.	.	.	.	.	r
Cryptolithoides sp.	.	.	.	.	.	r	.	.	.	.	.	.	.	.	.	.
Encrinurus sp. cf. *E. cybeleformis* Raymond	.	.	.	.	u	.			.	.	.	.	.	.	.	.
Eomonorachis (?) sp.	.	.	.	.	.	.	.	r	.	.	.	.	.	.	.	.
Hemiarges sp. cf. *H. paulianus* (Clarke)	.	.	.	.	u	.	.	.	.	.	.	.	.	.	.	.
Homotelus sp.	.	.	.	.	.	.	.	.	.	u	.	.	.	.	r	r
Illaenus sp. cf. *I. americanus* (Conrad)	.	.	.	.	.	.	c	.	.	.	.	.	.	.	.	.
Illaenus spp.	.	u	.	.	u	c	u	c	c	.	.	.	.	.	.	.
Thaleops sp.	.	.	.	.	.	.	r	u	r	.	.	.	.	.	.	.
Aparchites sp.	.	.	.	.	.	.	.	r	.	.	.	.	.	.	.	.
Eurychilina sp. cf. *E. reticulata* Ulrich	.	.	.	.	.	.	.	r	.	.	.	.	.	.	.	.
Leperditia sp.	.	.	.	.	.	.		u	.	.	.	.		.	.	.

* a, abundant; c, common; u, uncommon; r, rare. Localities: 1, Nealmont formation, "north" quarry, Gilland farm in Blacklog Valley, 1.8 miles S.63°E. of Orbisonia; 2, Nealmont formation, Grove (Binkley and Ober) Quarry, Blacklog Valley, 1.8 miles S.54°E. of Orbisonia; 3, Doylestown Quarry, 1.5 miles S.47°W. of Doylesburg, Path Valley; 4, Dry Run (Binkley and Ober) Quarry, Path Valley; 5, Spring Run (Hammond) Quarry, 0.8 miles S.5°E. of Spring Run, Path Valley; 6, Composite, Willow Hill village and Pennsylvania Turnpike to north, Path Valley; 7, Fort McCord section, 4.8 miles N.20°E. of St. Thomas; 8, Small quarry, 1.3 miles S.18°W. of St. Thomas; 9, Brandt Church section, 2.7 miles S.13°W. of St. Thomas; 10, Rockdale section, along Cumberland Valley Railroad, 0.5 miles N.45°W. of Williamson; 11, U. S. Route 40 at Wilson, Md.; 12H, Housum member, Diller Church section, along Pennsylvania Turnpike, 2.9 miles N.45°E. of Newville; 12K, Kauffman member, same locality; 13K, Kauffman member, field and road cut exposures at Middle Spring; 14H, Housum member, field and railroad cut, 2.1 miles S.53°W. of Marion; 14K, Kauffman member, same locality.

Mercersburg. The lower occurrence is in the lower half of the formation and contains *Sowerbyella* sp. cf. *S. punctistriata* Mather, *Hesperorthis tricenaria* (Conrad), and *Öpikina* sp. cf. *Ö. minnesotensis* (Winchell) as the most prominent associates. The upper abundance is within the upper 15 feet of the Mercersburg and contains *Öpikina* sp. aff. *Ö. subtriangularis* Wilson, *Öpikina* sp. cf. *Ö. transitionalis* (Okulitch), *Öpikina* sp. cf. *Ö. wagneri* (Okulitch), and *Valcourea* sp. cf. *V. loricula* (Hall) as associates. Elements of the lower occurrence were observed in Blacklog Valley and in the western belts of the Cumberland Valley.

The Kauffman member is thought to correspond to a lower part of the uppermost 240 feet of beds described by Stose (1909, p. 8) at the type section. According to Stose these beds contain "a few fossils which indicate the *Christiania* fauna." Ulrich (Stose, 1909, p. 9) referred 280 feet of limestone to the *Christiania* zone. This thickness probably includes all of the Mercersburg as here defined. Bassler (1919, p. 146) subdivided the *Christiania* zone into 150 feet of Greencastle above and 100 feet of *Christiania* bed below. The Kauffman seems to correspond to part of both of these units. In the western belts of outcrop the Kauffman member corresponds roughly to Ulrich's fourth faunal or *Nidulites* zone.

Correlation.—The Kauffman member of the Mercersburg formation correlates directly with the upper part of the Nealmont formation in Blacklog Valley. Exact relations depend upon the accuracy of metabentonite correlation. If the metabentonite in the Nealmont of Blacklog Valley is Mercersburg metabontonite 2 as previously indicated, and if this is the upper metabentonite (N_2) of the Nealmont of central Pennsylvania (Kay, 1944, p. 103), the Kauffman corresponds to the Centre Hall and Rodman members of the Nealmont formation.

The presence or absence of Mercersburg equivalents in the Edinburg formation of Virginia seems dependent on the age of the St. Luke member at the top of the Edinburg. If the *Cryptophragmus*-bearing beds of south-central Pennsylvania are equivalent to the St. Luke as implied by Cooper and Cooper (1946, p. 57–58), Mercersburg equivalents must be absent in Virginia, for the St. Luke is directly overlain by the Oranda formation. However, if the uppermost Edinburg is lower Trentonian as suggested by Cooper and Cooper (1946, p. 86), Mercersburg equivalents are probably present.

ORANDA FORMATION

Name.—Stose (1909, p. 8) makes the first mention of an interval of black, thick-bedded, chalky-weathering limestone at the top of the Chambersburg formation. The unit was recognized only from the Marion section southward to Greencastle. Ulrich (*in* Stose, 1909, p. 9) regarded it as a distinguishable facies but included it as a subzone within the *Christiania* zone because of faunal similarity. Bassler (1919, p. 143–144) recognized these beds as far north as Chambersburg, noted the recurrence of *Echinosphaerites*, and named the unit the Greencastle bed. The writer (1941b, p. 1946) raised the Greencastle to formational value and restricted the unit to 63 feet of black massive limestone instead of the 150 feet designated by Bassler at the Marion section. In keeping with the original limits of the Chambersburg,

the *Sinuites* zone was included in the Greencastle formation, thereby placing the contact with the Martinsburg at a prominent lithologic change.

Continued use of the term Greencastle is now thought inadvisable. The original use appears to have been inconsistent. The base of the Greencastle bed as described at Marion is at an indefinite horizon in the middle of the Kauffman member of the Mercersburg. In the section at Greencastle (Bassler, 1919, p. 147), beds resembling Housum were assigned to the *Christiania* zone; a metabentonite with a thick interval of drab siltstone and succeeding limestone suggesting Mercersburg metabentonite 2 and the Kauffman member were assigned to the Greencastle bed. Thus, the base of the Greencastle bed seems to have been drawn at the base of the Kauffman member at Greencastle and in the middle of the Kauffman near Marion. The upper part of the Greencastle section is poorly exposed and may be structurally complicated. Excessive restriction of the Greencastle bed was necessary in defining the Greencastle formation. In the belief that the *Sinuites* zone overlies a marked hiatus and despite the sharp lithologic break that succeeds the zone in this part of Pennsylvania, Ulrich (1911, p. 328) removed the *Sinuites* zone to the Martinsburg formation. This assignment has been followed by almost all subsequent authors.

Cooper and Cooper (1946, p. 86) proposed the name Oranda formation for a "zone characterized by the occurrence of *Reuschella "edsoni"* (Bassler) and its associates, which overlies the Edinburg and directly underlies the *Sinuites* beds of the Martinsburg" in northern Virginia. Except for the *Sinuites* zone, the beds formerly assigned to the Greencastle formation agree faunally and stratigraphically with this definition. The term Oranda formation is therefore extended to these beds in south-central Pennsylvania and the *Sinuites* zone is referred to the base of the Martinsburg formation.

Lithologic characteristics.—Through most of south-central Pennsylvania the Oranda formation consists of dark-grey to black, fine-grained, slabby to massive, blue-grey to white-weathering limestone. However, at the easternmost outcrops in Franklin County, those in the vicinity of Marion, the Oranda consists of black, argillaceous, buff-weathering limestone with prominent shaly partings. The formation is consistently underlain by the Kauffman member of the Mercersburg or beds thought to be equivalent and is overlain by the *Sinuites* zone of the Martinsburg. Both contacts are minor disconformities. In many places the Oranda is difficult to distinguish from the Mercersburg but it is typically darker, heavier bedded, or finer grained and has a few feet of coarsely crystalline or semi-crystalline limestone at the base. At Middle Spring (section 26) and near Spring Run (section 6) the Mercersburg-Oranda contact is well exposed. At both places the contact is annealed and shows fine pitting in the uppermost bed of the Mercersburg.

A total of three metabentonites have been recognized in the Oranda of south-central Pennsylvania. The uppermost bed is quite persistent and is correlated by stratigraphic position with Rosenkrans' (1934, p. 115) Salona metabentonite 0 of central Pennsylvania. The two lower metabentonites are much less persistent. In the eastern belts the middle clay is known only in the Oakville section (section 25) and near Green Spring. The lower clay is known only at Middle Spring and at the Pennsylvania Turnpike near Newville (section 23). In the western belts, only one

clay bed has been found beneath Salona O, and this at only one locality. This clay is 5 feet below the O clay at the Africa section, 1.6 miles S.63°W. of Mercersburg (section 11).

Thickness.—The Oranda formation thins uniformly to the south and west from Carlisle. It is over 158 feet northeast of Plainfield (section 22) and is 149 feet on the Pennsylvania Turnpike northeast of Newville. Two miles southwest of Marion, the Oranda is 52 feet. However, the more argillaceous beds in the vicinity of Marion are only 20–30 feet thick. The formation was not identified in any section in Maryland because of poor exposure. If present, it is probably quite thin. Westward from Carlisle, the Oranda is between 45 and 58 feet thick northeast of Edenville (section 16) and reaches minima of 17 feet south of Fannettsburg (section 8) and 18 feet at Willow Hill in Path Valley (section 7). From Path Valley, the Oranda diverges slightly to 23 feet in Blacklog Valley (section 1).

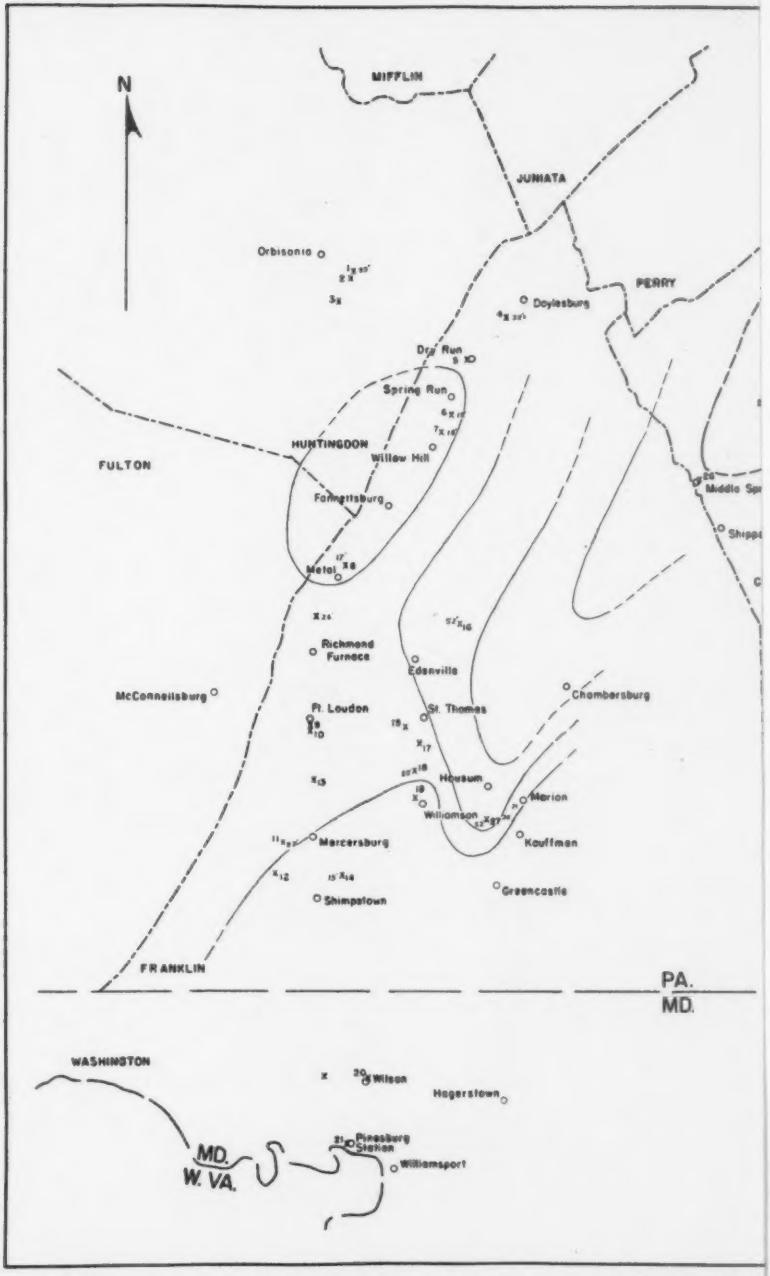
If the lower metabentonite at the Africa section correlates with one of the two lower clays found in Cumberland County, a southwestward convergence of either 28 or 134 feet of beds is indicated, depending on which metabentonite is correlated. Evidence of overlap or possibly of additional convergence in the lower beds of the Oranda is indicated by the position of the basal calcarenite with respect to Salona 0. Two miles southwest of Marion the base of the calcarenite is 34 feet below clay 0; west of Mercersburg it is 15 feet below; in northern Path Valley it is about 16 feet below; and in Blacklog Valley (sections 1 and 2) metabentonite 0 is within the basal calcarenite, only 1 foot above the base of the formation. Contrary to the westward convergence of the interval below Salona 0, the Oranda beds above clay 0 diverge from 3 feet in Path Valley to 22 feet in Blacklog Valley. The nature of this thickening is not known. The base of the *Sinuites* zone is not well marked in Blacklog Valley and the base of Salona metabentonite 2 was arbitrarily chosen as the top of beds equivalent to the Oranda. Part or all of the thickening may be the result of differential preservation of beds beneath a pre-*Sinuites* zone disconformity. However, a large part of the thickening may be true divergence of beds.

Isopachs of the Oranda formation (Pl. 5) indicate the northeasterly thickening from Marion to the vicinity of Carlisle and the westerly thinning to Path Valley.

Fossils.—The Oranda is quite fossiliferous (Table 8) and the fauna is similar to that of the Oranda in Virginia. *Reuschella edsoni* (Bassler), *Christiania trentonensis* Ruedemann, and *Birmuria lamellosa* (Bassler) are more common in the eastern belts of outcrop. *Receptaculites* sp. is more common in the western belts. *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens) and *Oxoplecia simulatrix* (Bassler) are common in both eastern and western belts. *Echinosphaerites* is present locally in the basal calcarenite and directly succeeding beds and represents the upper occurrence compared by Raymond (1916, p. 239) with unit 4bα of Norway.

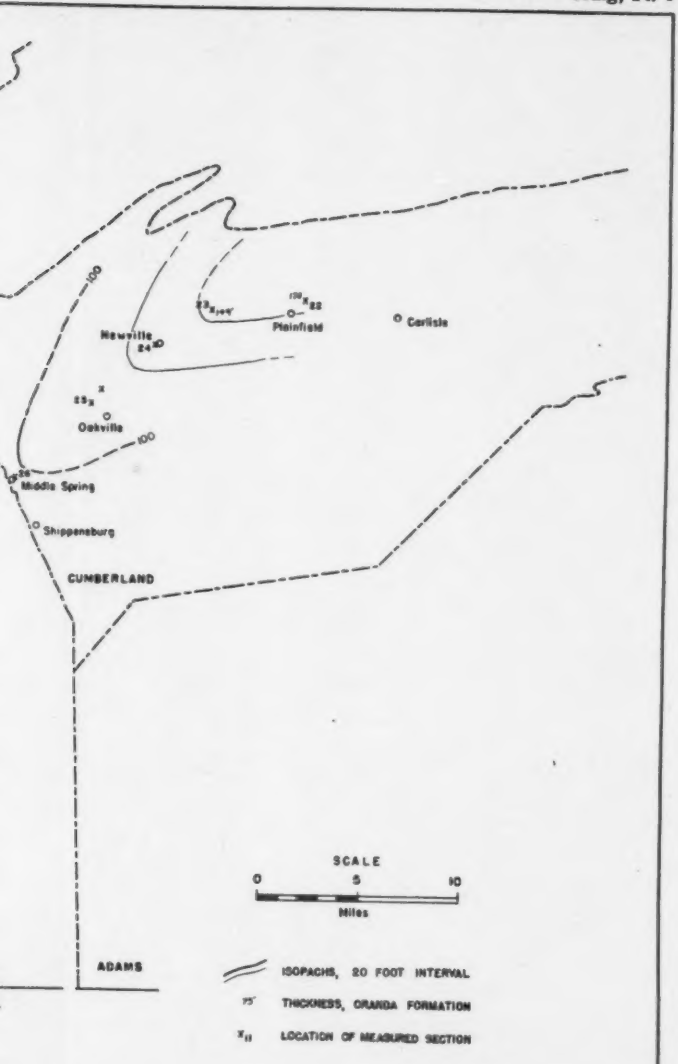
Correlation.—The Oranda formation correlates approximately with the part of the Salona formation below Salona metabentonite 2 in Blacklog Valley.

The upper part of the Jacksonburg limestone in New Jersey (Weller, 1903, p. 16–49; Miller, 1937, p. 1695) shows some faunal similarity to the Oranda and the *Sinuites* zone of the lowermost Martinsburg. Both the Oranda and upper Jacksonburg contain the lowest occurrences of *Prasopora orientalis* Ulrich and *Cryptolithus*



ISOPACH MAP OF THE ORANDA FORMATION

THE ORANDA FORMATION,



TION, SOUTH-CENTRAL PENNSYLVANIA

TABLE 8.—*Faunules from the Oranda formation**

	Localities									
	1	2	3	4	5	6	7	8	9	10
Receptaculites sp.	c	.	c	c	.	u	u	u	.	.
Echinosphaerites sp.	.	.	.	r	.	.	.	.	.	a
Prasopora sp. cf. *P. orientalis* Ulrich	.	.	a	.	c	a	c	a	a	.
Subretopora (*Chasmatopora*) sp.	.	.	.	.	.	.	r	.	.	.
Bimuria lamellosa (Bassler)	.	.	r	.	.	.	.	.	.	a
Christiania trentonensis Ruedemann	.	.	.	.	u	.	.	.	.	c
"Dalmanella" sp.	.	.	.	.	.	.	u	.	.	.
Dinorthis sp. cf. *D. pectinella* (Conrad)	.	.	.	.	.	r	.	.	.	.
Eoplectodonta sp. cf. *E. alternata* Butts	.	.	.		.	c	.	u	.	u
Glassia (?) sp.	.	.	.	.	.	u	.	.	.	.
Hesperorthis tricenaria (Conrad)	.	.	.	r	.	.		.	.	.
Leptaena sp. aff. *L. rhomboidalis* (Wilckens)	a	u	c	u	u	c	c	u	u	u
Leptaena sp.	.	.	.	.	.	.	.	u	u	.
Lingula sp.	.	c	.	.	.	.	r	u	c	.
Lingulasma sp.	u	.	.	.	.	.	.	.	.	.
Oxoplecia simulatrix (Bassler)	.	.	u	c	a	a	a	c	u	u
Reuschella edsoni (Bassler)	.	c	.	.	.	c	u	.	u	u
Schizocrania filosa (Hall)	.	.	.	.	.	.	.	.	.	r
Skenidioides (?) sp.	.	c	.	c	.	r	.	.	.	.
"Sowerbyella" sp. cf. *"S." pisum* (Ruedemann)	c	.	.	r	.	.	c	.	.	.
Sowerbyella sp. cf. *S. rugosa triradiatus* Butts	.	.	.	.	.	.	.	r	.	r
Sowerbyella sp.	c	c	a	c	u	c	u	c	c	u
Strophomena sp. aff. *S. planodorsata* Winchell and Schuchert	.	.	.	.	.	a	.	.	.	.
Strophomena sp. aff. *S. sculpturata* Bassler	.	.	.	r	.	.	.	.	.	.
Lophospira sp.	.	.	.	.	r	.	r	r	.	.
Phragmolites sp.	.	.	.	.	.	.	r	.	.	.
Cyrtoceras (?) sp.	.	.	.	.	.	r		.	.	.
Orthoceras sp. cf. *O. arcuoliratum* Hall	.	.	.	.	.	.	r	.	.	.
Spyroceras sp.	.	.	.	.	.	.	r	.	.	.
Bumastus sp.	.	.	.	.	.	.	.	u	.	.
Calliops sp. aff. *C. armatus* Ulrich and Delo	.	.	.	.	.	.	r	.	.	.
Calliops sp. cf. *C. callicephalus* (Hall)	.	.	.	.	u	c	c	u	.	.
"Calymene" sp.	.	.	.	.	.	r	r	.	.	.
Ceraurus sp.	.	.	.	.	.	.	u	.	.	.
Cryptolithus tesselatus Green	.	.	c	c	u	u	a	u	c	u
Encrinurus sp.	.	.	.	.	.	.	c	.	.	c
Eomonorachis sp.	.	.	.	.	.	r	.	r	.	u
Homotelus sp.	.	.	.	.	.	c	.	.	u	.
Illaenus sp.	.	.	.	.	u	u	.	r	.	.
Tetralichas trentonensis (Conrad)	.	.	.	.	.	u	.	.	.	.

* a, abundant; c, common; u, uncommon; r, rare. Localities: 1, Doylestown Quarry, 1.5 miles S.47°W. of Doylesburg, Path Valley; 2, Africa section, 1.6 miles S.63°W. of Mercersburg; 3, Fort McCord section, 4.8 miles N.20°E. of St. Thomas; 4, Brandt Church section, 2.7 miles S.13°W. of St. Thomas; 5, Elliotts Mill section, 0.9 miles N.42°E. of Plainfield; 6, Logan School, 2.6 miles N.78°W. of Plainfield; 7, Pennsylvania Turnpike, 2.9 miles N.45°E. of Newville; 8, Green Spring Quarry, 1.3 miles N.12°W. of Oakville; 9, Oakville section, 1 mile N.52°W. of Oakville; 10, Marion section, field exposure near Cumberland Valley Railroad, 2.1 miles S.53°W. of Marion.

tesselatus Green. Other forms contained in common are *Tetralichas trentonensis* (Conrad), *Calliops callicephalus* (Hall), *Calymene* sp. and *Encrinurus* sp. Partial equivalence of the Oranda formation and upper Jacksonburg limestone is indicated.

MARTINSBURG SHALE

Name.—The Martinsburg shale (Geiger and Keith, 1891, p. 161), named for Martinsburg, West Virginia, has been variously estimated as 2000 to 2500 feet of shale and sandstone in south central Pennsylvania and Maryland. Only the lower few hundred feet of the formation have been included in the present study. The Martinsburg succeeds the Oranda formation in south-central Pennsylvania.

Lithologic characteristics.—The *Sinuites* zone consists of black, muddy, rubbly to slabby limestone and forms the thin basal unit of the Martinsburg throughout the area. Evidence of a minor disconformity between the Oranda and the *Sinuites* zone has been observed in most sections southeast of Blacklog Valley. In Path Valley several sections exhibit a bed of limestone conglomerate with pebbles up to 3 inches in diameter at the base of the *Sinuites* zone. In the Cumberland Valley, the base of the zone is coarsely crystalline to semicrystalline limestone. In the quarry at Green Spring between Oakville and Newville in Cumberland County, the basal beds of the *Sinuites* zone lens and locally overlap on the Oranda formation.

A sharp lithologic change succeeds the *Sinuites* zone throughout most of the area. In the easternmost belts of outcrop, those extending southward from Chambersburg, the overlying beds are black, platy to fissile, calcareous shale. In the Shippensburg-Carlisle belt and in all the western belts of outcrop, these beds are black, slabby to massive, argillaceous, buff-weathering limestone. Where highly fractured, this limestone weathers to a shaly debris. In Blacklog Valley the post-*Sinuites* zone beds have a less argillaceous aspect. From east to west the lower part of the Martinsburg exhibits a gradational change from more argillaceous to more calcareous beds.

Metabentonite beds are numerous in this part of the Martinsburg (Fig. 3). Eighteen separate beds have been found in the lower 220 feet of the formation at the composite Africa section, 1½ miles west of Mercersburg (section 11). Bonine and Honess (1929), Whitcomb (1932) and Rosenkrans (1934) have studied and correlated the metabentonites of the Salona in central Pennsylvania. Kay (1944, p. 110) has extended the correlation from central Pennsylvania to Path Valley. Eight clays have been recognized and numbered from 0 at the bottom to 7 at the top. Correlation to the south and east of Path Valley is difficult because of the increased number of clays and the loss of distinguishing criteria. Nevertheless, an attempt has been made to extend the Salona metabentonites through south central Pennsylvania.

Metabentonite 0 has been discussed in connection with the Oranda formation. Salona metabentonite 2 is the most persistent and readily correlated metabentonite in south-central Pennsylvania. The constant association of this clay with the *Sinuites* zone affords both lithologic and faunal bases of correlation. Metabentonite 1, which in central Pennsylvania is 2 feet beneath clay 2, has been identified in only two of the sections of south-central Pennsylvania. Two miles north of Richmond Furnace, clays 1 and 2 are separated by less than one-half foot of blocky shale. At

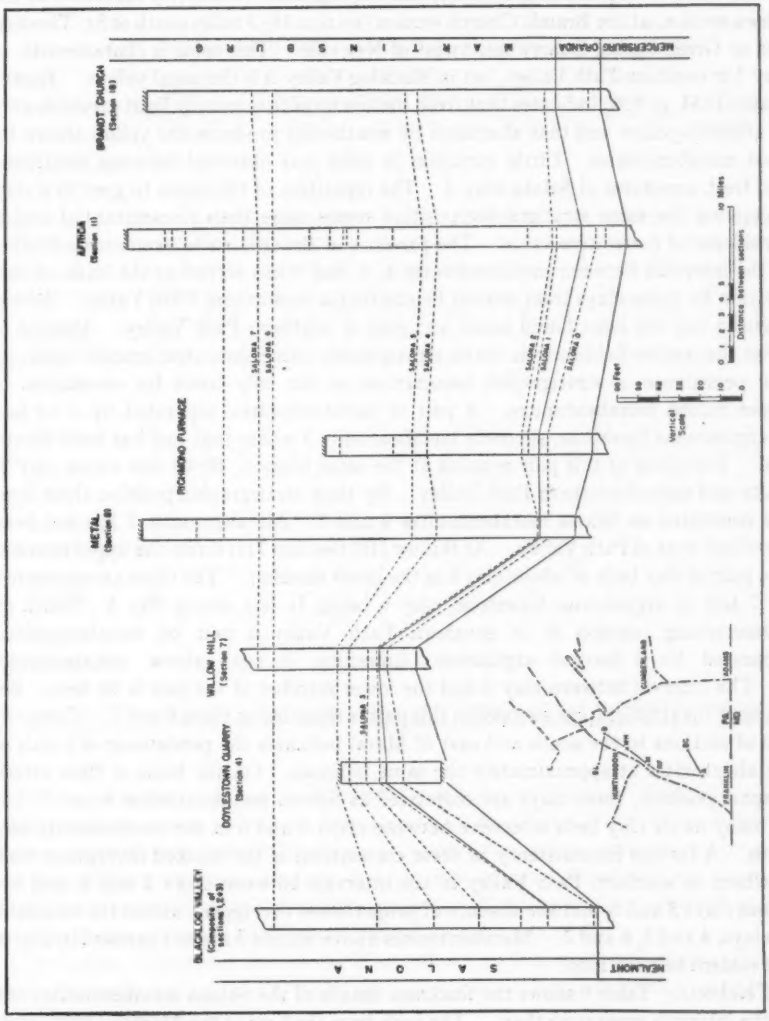


FIGURE 3.—*Correlation of Salona metabentonites in south-central Pennsylvania*

the Africa section (section 11) they are separated by 1.9 feet of black, argillaceous limestone. Salona metabentonite 3 is not persistent south and east of Path Valley. On the basis of its green to grey color, this clay has been tentatively identified at the Africa section, at the Brandt Church section (section 18) 3 miles south of St. Thomas, and at Green Spring Quarry southwest of Newville. This color is characteristic of clay 3 in northern Path Valley, but in Blacklog Valley it is the usual yellow. Rosenkrans (1934, p. 126) indicates that fresh metabentonite is usually light greenish-grey or greenish-yellow and that alteration by weathering produces the yellow shown by most metabentonites. Little variation in color was observed between weathered and fresh exposures of Salona clay 3. The repetition of the green to grey in a clay occupying the same stratigraphic position seems more than circumstantial and is considered of correlative value. The presence of *Brongniartella trentonensis* (Collie) in the intervals between metabentonites 4, 5, and 6 has served as the basis of correlation for these clays from central Pennsylvania to northern Path Valley. *Brongniartella* has not been found south and east of northern Path Valley. Absence of other distinctive field criteria leave stratigraphic succession, stratigraphic position, and persistence of stratigraphic associations as the only bases for correlation of upper Salona metabentonites. A pair of metabentonites, separated by 5–12 feet of argillaceous limestone, succeeds metabentonite 3 where that bed has been identified. The lower of this pair remains at the same horizon, 50–60 feet above clay 2, south and east of northern Path Valley. By their stratigraphic position these beds are correlated as Salona metabentonites 4 and 5. Metabentonite 7 has not been identified west of Path Valley. At Willow Hill (section 7) it forms the upper member of a pair of clay beds of which clay 6 is the lower member. The clays are separated by 7 feet of argillaceous limestone, clay 6 being 31 feet above clay 5. South of Fannettsburg (section 8) in southern Path Valley a pair of metabentonites, separated by 6 feet of argillaceous limestone, is next above metabentonite 5. The interval between clay 5 and the lower member of the pair is 66 feet. Because of the stratigraphic succession this pair is regarded as clays 6 and 7. Comparison of sections to the south and east of Metal indicates the persistence of a pair of metabentonites at approximately the same horizons. On the basis of their stratigraphic position, these clays are correlated as Salona metabentonites 6 and 7, but as many as six clay beds intervene between clays 5 and 6 in the southeasterly sections. A further inconsistency in these correlations is the marked divergence from northern to southern Path Valley in the intervals between clays 2 and 4, and between clays 5 and 6, and the absence of proportional divergence within the two pairs of clays, 4 and 5, 6 and 7. Metabentonites above Salona 3 are not exposed in any of the eastern belt sections.

Thickness.—Table 9 shows the thickness details of the Salona metabentonites and of the intervals separating them. The beds from the base of the Martinsburg formation to metabentonite 6 are 138 feet at the Brandt Church section (section 18) and 149 feet at the Africa section (section 11). In southern Path Valley this interval is 129 feet but converges abruptly to 51 feet at Willow Hill (section 7) in northern Path Valley. The interval diverges again to 133 feet in the Salona limestone of Blacklog Valley, and at Union Furnace in central Pennsylvania (Kay, 1944, p. 110) the minimum thickness of equivalent beds is 191 feet.

The few complete sections of the lower Martinsburg do not provide sufficient control for detailed isopachous study. However, the isopachs do cut across the western belts of outcrop and probably have a northeastward trend.

TABLE 9.—*Comparison of Salona metabentonites in south-central Pennsylvania**

Horizon or unit interval	Blacklog Valley composite (sections 1, 2, 3)			Willow Hill, Path Valley (section 7)			Metal, Path Valley (section 8)			North of Richmond Furnace, Path Valley			Africa (section 11)			Brandt Church (section 18)		
	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3	1	2	3
Clay 7	—	—	—	—	0.2	58	—	0.2	3	—	—	—	—	0.8	181	—	0.1	169
Interval	—	—	—	7	—	—	6	—	—	—	—	—	7	—	—	10	—	—
Clay 6	—	0.3	155	—	0.6	51	—	0.3	146	—	—	—	—	0.6	173	—	0.2	159
Interval	48B	—	—	31B	—	—	66	—	—	—	—	—	72	—	—	65	—	—
Clay 5	—	1.6	107	—	0.4	19	—	1.0	80	—	0.4	87	—	0.5	100	—	0.2	94
Interval	26B	—	—	5B	—	—	12	—	—	7	—	—	7	—	—	5	—	—
Clay 4	—	0.9	79	—	0.3	14	—	1.0	67	—	0.2	80	—	0.5	92	—	0.8	89
Interval	21	—	—	8	—	—	—	—	—	—	—	—	43	—	—	29	—	—
Clay 3	—	1.5	57	—	0.5	6	—	—	—	—	—	—	—	0.4	48	—	0.5	59
Interval	30	—	—	3	—	—	48	—	—	45	—	—	12	—	—	31	—	—
Clay 2	—	0.7	25	—	0.7	2	—	1.0	19	—	1.9	35	—	0.4	36	—	0.9	27
Interval	—	—	—	—	—	—	—	—	—	0.4	—	—	1.9	—	—	—	—	—
Clay 1	—	—	—	—	—	—	—	—	—	—	1.1	33	—	0.1	34	—	—	—
Interval	0	—	—	1	—	—	1	—	—	5	—	—	12	—	—	5	—	—
Base of Martinsburg	—	—	24	—	—	18	—	—	17	—	—	27	—	—	22	—	—	21
Interval	22	—	—	4	—	—	17	—	—	4	—	—	6	—	—	15	—	—
Clay 0	—	0.3	2	—	0.3	14	—	—	—	—	0.5	23	—	0.7	16	—	?0.3	6
Interval	2	—	—	14	—	—	—	—	—	22	—	—	15	—	—	6	—	—
Base of Oranda	—	—	0	—	—	0	—	—	0	—	—	0	—	—	0	—	—	0

* Figures in columns 1 and 2 are individual thicknesses of units; figures in column 3 are cumulative thicknesses to base of Oranda formation. All measurements in feet. B after thicknesses indicates presence of *Brongniartella* in interval.

Fossils.—The lower part of the Martinsburg is quite fossiliferous (Table 10). The lowest beds are characterized by *Sinuites cancellatus* (Hall), *Cryptolithus tesselatus* Green and *Leptobolus ? ovalis* Bassler. The succeeding more argillaceous beds contain *Cryptolithus tesselatus* Green, *Climacograptus* sp., and *Corynoides* sp. *Brongniartella trentonensis* (Collie) is found in Path and Blacklog Valleys where it is apparently restricted to the beds between metabentonites 4 and 6.

Correlation.—The lower part of the Martinsburg formation is equivalent in Blacklog Valley to the part of the Salona limestone succeeding metabentonites 1 and 2 and including metabentonite 6.

The presence of *Sinuites cancellatus* (Hall) and *Ecculiomphalus trentonensis* (Conrad) in the upper part of the type Jacksonburg limestone in New Jersey (Weller, 1903, p. 16–49) suggests that it may contain beds corresponding to the *Sinuites* zone of south-central Pennsylvania. Whether the Jacksonburg contains equivalents of beds younger than the *Sinuites* zone is not known. R. L. Miller (1937, p. 1701–1702, 1708–1709) has shown that the Martinsburg overlies the Jacksonburg disconform-

ably in New Jersey and eastern Pennsylvania and the age of the lowest beds of the Martinsburg is unknown in these areas.

Rosenkrans (1933) has tentatively identified Salona metabentonites 1, 2, and 3 southward to the vicinity of New Market, Virginia. The similarity of the associa-

TABLE 10.—*Faunules from the lower part of the Martinsburg shale**

	Localities					
	1	2	3	4	5	6
Climacograptus sp.	a	c	.	.	c	c
Corynoides sp.	u	.	.	.	.	c
Prasopora sp. cf. *P. orientalis* Ulrich	.	.	.	u	.	.
"Dalmanella" sp.	u	c	.	.	.	.
Leptaena sp.	.	.	.	c	.	.
Leptobolus ? ovalis Bassler	c	u	a	.	c	c
Lingula sp. cf. *L. riciniformis* (Hall)	.	.	.	.	.	u
Lingula sp.	u	u	.	.	.	u
Orbiculoidea sp. cf. *O. lamellosa* (Hall)	r	u	.	.	.	.
Strophomena sp. aff. *S. sculpturata* Bassler	u	c	.	u	c	.
Ecculiomphalus trentonensis (Conrad)	u	.	r	.	.	.
Sinuites cancellatus (Hall)	c	.	.	.	c	.
Brongniartella trentonensis (Collie)	u	.	.	.	.	.
Calliops sp.	.	.	.	.	.	u
Calymene sp.	.	.	u	.	.	.
Cryptolithus tesselatus Green	a	.	a	a	c	a
Encrinurus sp.	.	.	c	.	.	.
Homotelus sp.	.	.	.	c	c	.
Triarthrus becki (Green)	.	.	.	.	.	a

* a, abundant; c, common; u, uncommon; r, rare. Localities: 1, Pennsylvania Turnpike, north of Willow Hill, Path Valley; 2, Brandt Church section, 2.7 miles S.13°W. of St. Thomas; 3, Diller Church section, along Pennsylvania Turnpike, 2.9 miles N.45°E. of Newville; 4, Green Spring Quarry, 1.3 miles N.12° W. of Oakville; 5, Oakville section, 1 mile N.52°W. of Oakville; 6, Roadcut west of U. S. Route 11 in southern environs of Marion.

tion of metabentonites 1 and 2 to their association in Pennsylvania combined with their occurrence immediately above the *Christiania*-bearing Oranda verifies the correlation of these beds. The identification of metabentonites 0, 3, 4, and 5 in Virginia seems less certain.

CLASSIFICATION

The Middle Ordovician of the Appalachians should be most satisfactorily correlated with the standard column of New York by the route of most continuous outcrop, that through eastern Pennsylvania, New Jersey, and southeastern New York. However, three factors make this route unsatisfactory: (1) the incomplete character of the Middle Ordovician of eastern Pennsylvania and New Jersey, (2) the inexact knowledge of its relation to the south-central Pennsylvania sequence,

and (3) the short discontinuities because of coverage by the Taconic klippe in New York and the more recently recognized thrust outlier proposed by Kay (1941c) in Pennsylvania and elaborated as the Hamburg klippe by Stose (1946, p. 665–696). The alternate route is by way of the northernmost anticlines of central Pennsylvania. In this area at least part of the Middle Ordovician section is more complete than the section in south-central Pennsylvania. The gap between the upper Mohawk Valley in New York and Mosquito Valley, the northernmost anticline exposing Middle Ordovician in central Pennsylvania, is about 170 miles.

Kay (1944, p. 111–112) has correlated the Salona limestone with the Shoreham and Denmark formations of the type locality of the Trentonian in New York. The beds below Salona metabentonite 1 are classified as Shoreham and those above as Denmark. Extending the correlation to south-central Pennsylvania, the lower part of the Martinsburg is equivalent to the Denmark. In the western belts of outcrop, the lower part of the Martinsburg is similar to the Denmark formation. Both are predominantly dark-grey to black, fine-grained, thick-bedded limestones, although, the lower Martinsburg is more argillaceous than the Denmark. *Sinuites cancellatus* (Hall), common at the base of the Martinsburg, is abundant in the *Trocholites* zone at the base of the Poland member of the Denmark (Craig, 1941a, Kay, 1943b, p. 600). A thin metabentonite overlies the *Trocholites* zone in New York, a position similar to that of Salona metabentonite 2 with respect to the *Sinuites* zone. Succeeding this clay, four metabentonites have been recognizd in the Denmark and Kay has compared them with Salona metabentonites 3 to 6 on the basis of stratigraphic succession.

The Oranda formation, dark, fine-grained, thick-bedded limestone, is correlated with the Shoreham, thin-bedded calcarenite and shale. Both contain the first appearance of *Cryptolithus tesselatus* Green and *Prasopora orientalis* Ulrich. Both contain *Reuschella edsoni* (Bassler). Although one metabentonite may be present locally in the Shoreham, it has not been correlated to south-central Pennsylvania.

The Mercersburg formation is classified as Rockland and Kirkfield, the lower formations of the Trentonian in New York. The Mercersburg overlies a regional unconformity and underlies beds equivalent to the Shoreham, a stratigraphic position identical with that of the Rockland and Kirkfield limestones. Although the faunal sequence found in New York has not been recognized, two of the more distinctive Rockland and Kirkfield species are found in the Mercersburg. These are *Encrinurus cybeleformis* Raymond and *Hemiarges paulianus* (Clarke). Species of *Triplesia*, *Gonioceras*, and *Phragmolites* in the Mercersburg may support this correlation.

Classification of the Shippensburg formation is more difficult. Kay has presented two interpretations of the classification of the Benner formation of central Pennsylvania. By the earlier interpretation (Kay, 1944, p. 22–23) the Snyder and Stover members of the Benner were correlated with the Pamelia. This was based on the abundant occurrence of *Cryptophragmus* in the Pamelia of Ontario and the correlation of metabentonite A near the top of the Valley View member of the Curtin formation with a single metabentonite in the base of the Lowville of central New York. Subsequently Kay (1946, p. 1210) has tentatively correlated the Snyder with the Lowville and the Stover with the Chaumont, basing it on the persistence of the Sny-

der and Stover lithic types from Virginia to central Pennsylvania and their close similarity to the Lowville and Chaumont in New York. Despite the apparent conflict with faunal evidence, the latter correlation seems preferable. Like the Snyder, the calcilutite facies of the Doylesburg member is similar to the Lowville. Both contain *Tetradium cellulosum* (Hall), *T. fibratum* Safford, and *Cryptophragmus antiquatus* Raymond. It may be noted that the abundant occurrence of *Cryptophragmas* in the upper Pamelia at Cobokonk, Ontario, does not persist throughout Ontario and New York. Young (1943, p. 233) has recorded only two occurrences of *Cryptophragmus* in the Black River of northwestern New York and southeastern Ontario. Both were rare occurrences and neither were in the upper Pamelia.

If the Doylesburg is Lowville in age, the underlying Fannettsburg member contains Pamelia equivalents at least in the upper part. The rare presence of *Cryptophragmus* in the upper Fannettsburg is compatible with this correlation. By this interpretation the Pinesburg and lower Fannettsburg may be either Pamelia or may be partly of pre-Pamelia age.

Classification of the beds between the Beekmantown and the Pinesburg has varied. In part it has been incorrect as a result of misinterpretation of the stratigraphic sequence within south-central Pennsylvania. Ulrich (Stose, 1910, p. 8) regarded the middle part of the "Stones River" as faunally equivalent to the middle Chazyan. Cooper and Cooper (1946, p. 70, 74) indicate the partial or complete equivalence of Whistle Creek limestone of Virginia to the middle "Stones River." The Whistle Creek is regarded as probably younger than the middle of the type Lenoir which contains fossils of the Crown Point and Valcour divisions of the Chazyan. Thus, the middle "Stones River" could be no older than upper Valcour, if Chazyan at all, and the upper "Stones River" is at least post-Crown Point and may be post-Chazyan.

Oddly enough, the upper "Stones River" contains *Tetradium cellulosum* (Hall) and other species suggesting the Black River of New York. The succeeding Pinesburg and Fannettsburg have both Black River and Chazyan faunal aspects. Species of *Mimella*, *Multicostella*, and *Camerella* suggest the Chazyan. Species of *Öpikina* and *Strophomena* suggest Black River age. The early appearance of a predominantly Black River fauna in the upper "Stones River" may be the result of facies control and the interval could be synchronous with an upper part of the Valcour. Thus, the Pinesburg occupies a part of the interval between middle Chazyan and upper Pamelia. It may either be entirely of Pamelia age or may partially complete the Chazy-Black River hiatus in the New York column.

Kay (1947, p. 1198) has proposed the Bolarian group to include all beds of post-late Chazyan and pre-early Trentonian age. The Bolarian is divided into two subgroups, the Hatterian below and Hunterian above. The Hatterian in central Pennsylvania is represented by the Eyer, Grazier, and Hostler members of the Hatter formation. In south-central Pennsylvania the Pinesburg and Fannettsburg are Hatterian. The Hunterian in central Pennsylvania includes the Snyder and Stover members of the Benner formation and the Curtin limestone. In south-central Pennsylvania the Doylesburg is the only representative of the Hunterian.

STRUCTURAL CONSIDERATIONS

The sharpest deflection in the rather sinuous course of the Appalachian Valley is in Franklin and Cumberland Counties, Pennsylvania. In the 40 miles between Greencastle and Carlisle, the valley changes course 50 degrees from the N.20°E. trend in Maryland, West Virginia and northern Virginia to the N.70°E. trend east of Harrisburg, Pennsylvania. Variations in deformation along the arc of this deflection appear slight. As shown by existing geologic maps, a somewhat greater amount of reverse faulting and somewhat closer folding in southern Franklin County suggest greater deformation than to the north and east. This is thought to be more apparent than real. Ernst Cloos (1940) has discussed the bulge in the Appalachians of Pennsylvania and presented data showing the increase of crustal shortening toward Harrisburg from northeastern Pennsylvania. Southward from the bulge into Maryland, a decrease in crustal shortening may be expected. The overall shortening of at least the surficial beds is considerable, but no conclusive field evidence indicates differential crustal shortening in the area.

The crowding of isopachs of the Shippensburg and Mercersburg between Marion and Williamson suggests that the eastern belts have an excessive westward displacement. This would help explain the contrast between the eastern and western sequences of the lower Middle Ordovician. However, isopach studies of the members indicate that this is probably an original stratigraphic contract, rather than an imposed structural contrast. Isopachs of the Pinesburg member of the Shippensburg show no crowding between Marion and Williamson, but a uniform westward thinning. Furthermore, a bottom slope of only 1°3′ between Williamson and Marion could have accounted for the excess thickness in the "Stones River"-Martinsburg interval of the eastern belts.

The transgression of isopachs and transgression of sections of similar character across the present belts of outcrop indicate that the present structures are not parallel to the structures of the lower Middle Ordovician. Except for a few minor details, southern sections in a westerly position are similar to northern sections with a more easterly position.

Variations in the thickness of units cannot be explained by flowage of beds during orogeny. If this were the case, the essentially random location of sections measured in this study should have produced a random distribution of thickness rather than the definite pattern of thickness variations actually observed.

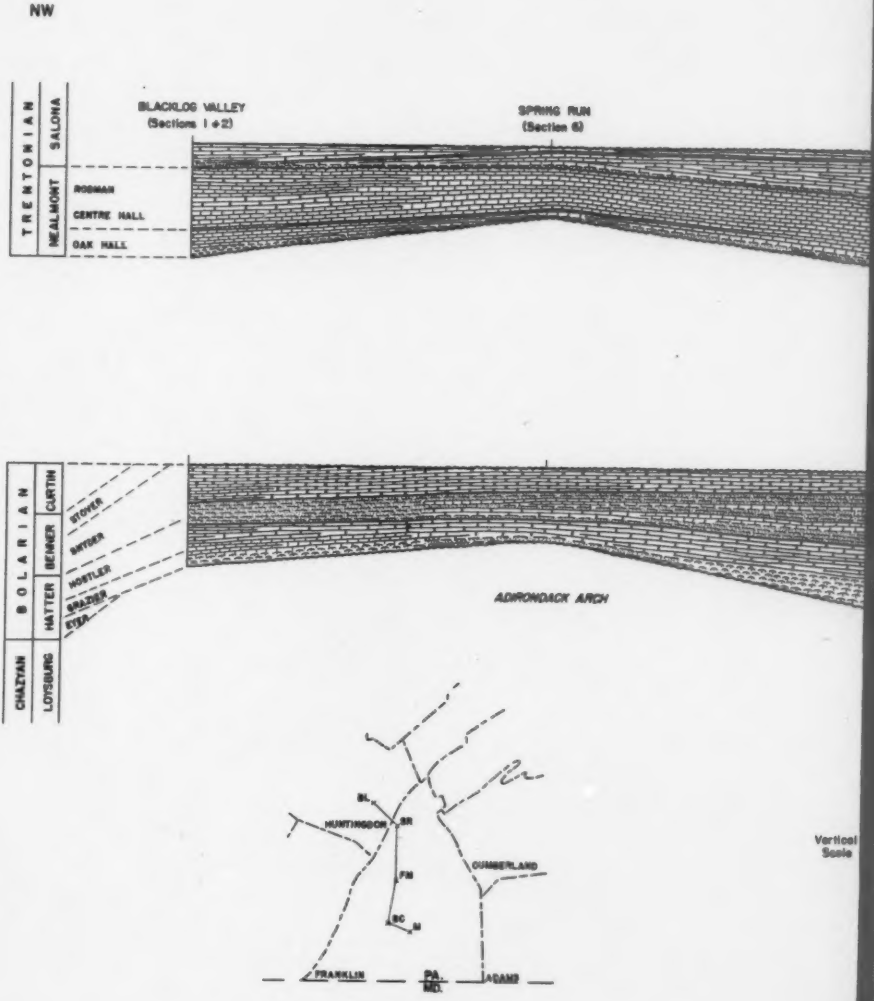
PALEOGEOGRAPHIC INTERPRETATIONS

Thickness studies indicate the persistence of a relatively positive structural area throughout the deposition of the described beds. Although the structure was positive relative to adjacent areas, the fact that most of the described units were deposited across the summit indicates that behavior was actually somewhat negative. The summit of the area is near Willow Hill and Spring Run in northern Path Valley (Pl. 6). The extension of the summit cannot be determined from exposure. However, northeastward-trending isopachs lateral to the high structure indicate that it

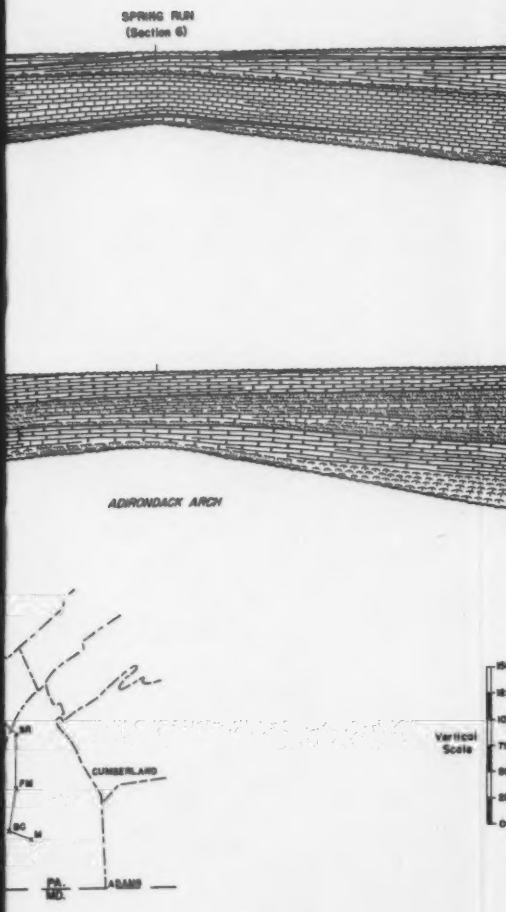
was linear in character. The general similarity of stratigraphic behavior and the close approximation of position by projection of isopachs indicate that this high was along the southward extension of the Adirondack line of New York. The thick sections east of the line were deposited in the Champlain miogeosyncline. The thick units west of the line were deposited in the Rideauan trough (Kay, 1944, p. 99) and in the Allegheny belt (Kay, 1942b, p. 1605) of central Pennsylvania.

Shippensburg deposition began with the westward marine overlap of the Pinesburg member toward and finally transgressing the Adirondack line. A similar overlap of the Hostler and Grazier on the Eyer took place from the west. Epeirogenic depression, producing the Adirondack arch, combined with some eustatic rise were required to produce this sequence. The conditions of deposition of the Fannettsburg were quite different. The Adirondack arch remained as a relatively positive feature but sloped only gradually eastward to a flexure marking the western limit of the Champlain belt and located between the eastern and western belts in the southern part of Franklin County. The flexure was formed by continued epeirogenic downwarp on the east which permitted the accumulation of the thick normal marine limestone of the Fannettsburg southwest of Marion. West of the flexure shallow-water deposition is reflected in the coarse, laminated coquinoidal character of the Fannettsburg. Metabentonites were not preserved in this area and the ash that fell may have been redistributed to deeper water, possibly accounting for the characteristic impure limestones succeeding Shippensburg clays 4 and 6 in the eastern sections. Isopachs of the Doylesburg member are quite irregular and reflect the removal of beds during the post-Shippensburg hiatus. The dove calcilutite was probably a shallow-water deposit. The few genera and species found in this facies indicate unfavorable environment, perhaps resulting from soft and suffocating bottom conditions. The dark, fine-grained limestones indicate a more normal marine environment and the limestone-pellet conglomerates show reworking of previously deposited and partially lithified beds. The predominence of these at the margin of the present limit of the Doylesburg suggests that the Doylesburg was originally deposited farther east, that it was probably a normal marine limestone formed in somewhat deeper water, and that these eastern beds were removed during the post-Shippensburg hiatus. The conglomerates may be the result of wave and current action at the eastern margin of the shallow platform. If this interpretation is correct, the structural pattern during the Doylesburg deposition was quite similar to that during the Fannettsburg: a shallow-water area through most of the western belts, a marginal epeirogenic flexure in approximately the previous position, and a deeper-water area to the east.

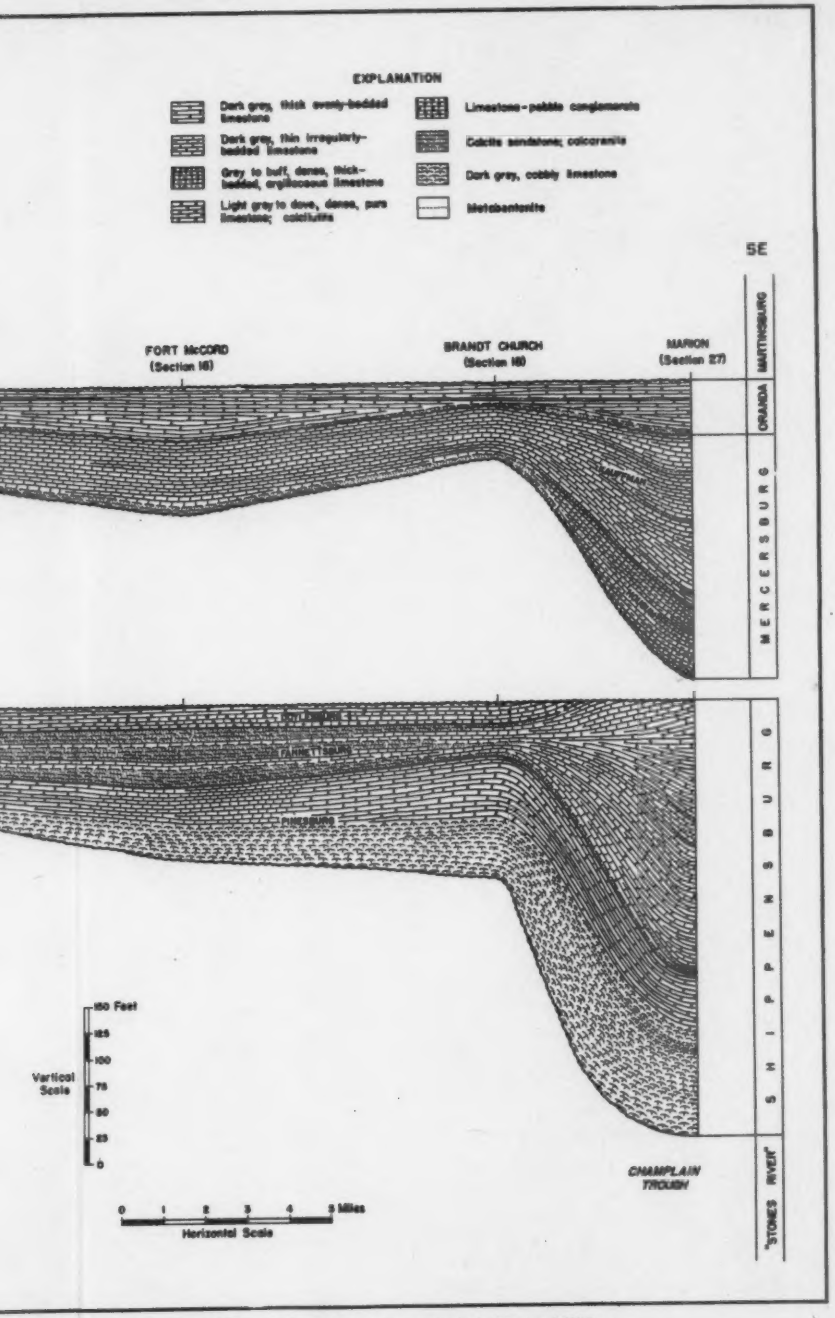
Beds equivalent to the Curtin formation and the Stover member of the Benner formation are not recognized in the Chambersburg region. In central Pennsylvania these units complete a part of the post-Shippensburg-pre-Mercersburg hiatus of south-central Pennsylvania. The original extent of the units is unknown for within central Pennsylvania they are truncated southeastward to extinction by the regional unconformity (Kay, 1944, p. 115). The Doylesburg cannot represent a transcending facies in a sequence of eastward replacing overlap. Metabentonites representing time lines in the overlying Stover and Valley View limestones (Kay, 1944, p. 17–21). are successively truncated by the pre-Nealmont unconformity in central Pennsyl-



RESTORED SECTIONS OF BOLARIAN A
IN SOUTH-CENTRAL P



SECTIONS OF BOLARIAN ANI
IN SOUTH-CENTRAL PEN



N AND TRENTONIAN FORMATIONS
AL PENNSYLVANIA

vania. The metabentonites are not represented in the Doylesburg, which must be older by superposition.

Structural conditions similar to those of the Fannettsburg and Doylesburg prevailed during the deposition of the Mercersburg formation. The Adirondack line marked the position of a structural arch. Again, the thicker sections of the eastern belts indicate relatively greater depression than occurred in the western belts. Part of the westward thinning may be due to overlap by the Housum member of the Mercersburg. Except for the somewhat argillaceous facies of the Housum in the eastern belts and the basal calcarenite of the western belts, the Mercersburg appears to have been a normal marine limestone deposited at fair depth. The Mercersburg converges northeastward from Marion in the eastern belts of outcrop.

The Oranda formation consists mostly of marine limestone. The prevailing thick and even bedding indicates deposition under more quiescent conditions than those typical of the preceding formations, and may reflect a moderation in climate or greater depth of water. In contrast to the Mercersburg, the Oranda formation diverges northeastward from Marion. Westward thinning of the Oranda demonstrates the continued positive structural behavior along the Adirondack line. However, eastward divergence of the lower Oranda from Path Valley without comparable westward divergence indicates that it behaved as a monoclinal flexure with depression to the east. Behavior as an arch was resumed during the deposition of the upper Oranda. The abundant argillaceous material in the easternmost exposures of the Oranda just south of Marion represents the first influx of the detrital material that forms the bulk of the succeeding Ordovician. The thin character of these argillaceous sections suggest an approach to marginal conditions to the east.

The lower part of the Martinsburg shale grades from a quite argillaceous and shaly limestone in the eastern belts to a relatively pure limestone in Blacklog Valley, most of the change taking place between Path and Blacklog Valleys. Divergence of the lower Martinsburg to the southeast and northwest from northern Path Valley demonstrates the persistence of the Adirondack arch through this deposition. The arch may have been effective as a barrier, restricting most of the argillaceous material to the eastern side. The base of the shale facies of the Martinsburg transgresses westward from just above the metabentonite-bearing interval in the eastern belts of south-central Pennsylvania to the base of the Antes Gap shale of central Pennsylvania. Thus, the Coburn limestone and perhaps an upper part of the Salona limestone are represented by shale in south-central Pennsylvania.

The history of the region is similar to that of the West Canada Creek-Mohawk Valley region of New York, with the following stratigraphic contrasts: In Pennsylvania, sediments of Pamelia and pre-Pamelia age overlap to the Adirondack arch; in New York the oldest Middle Ordovician sediments adjacent to the Adirondack line are Pamelia in age and are restricted to the western side. Rockland beds are the first to definitely show the existence of a structural arch along the Adirondack line in the Mohawk Valley; thus, a structural arch is known earlier in Pennsylvania than in New York. The Chaumont limestone is present in the Champlain miogeosyncline in New York, but has not been recognized in the Champlain belt in Pennsylvania. The influx of argillaceous material into the Champlain miogeosyncline was in Shore-

ham time in Pennsylvania but not until Denmark time in New York. This may indicate an earlier inception of the Vermontian disturbance in Pennsylvania. No interbedded limestone and shale facies similar to the transgressional "Dolgeville" has been recognized in Pennsylvania. Instead, the shale of the Denmark part of the Martinsburg grades westward to less argillaceous beds.

SUMMARY

Lower Middle Ordovician limestones have been correlated through south-central Pennsylvania and Maryland. The correlation for Blacklog and Path Valleys and the western belts of outcrop in the Cumberland Valley is highly reliable. Abundant lithologic and faunal contrasts permit recognition of relatively thin intervals along these outcrops. Within the eastern belts of outcrop in the Cumberland Valley of Pennsylvania and in all the belts in Maryland, reliability is somewhat less. Despite the increased thickness of the beds in these belts and the loss of most of the contrasting facies, satisfactory correlations are provided by metabentonites and, to a lesser extent, lithologic characteristics and fauna. The contrast between the western and eastern belts makes this the least reliable point of the correlation. Progressive changes cannot be traced because of absence of intermediate exposures. The lower and upper parts of the sequence are correlated with assurance across the gap; the middle part is correlated on negative and a small amount of positive evidence, but represents the most satisfactory correlation possible with present information.

The Shippensburg and Mercersburg are redefined by removal of 67 feet of beds from the Shippensburg to the Mercersburg. The Greencastle formation is discarded in favor of the Oranda formation, the *Sinuites* zone being removed to the base of the Martinsburg shale. The Shippensburg formation is divided into three new members, the Pinesburg, Fannettsburg, and Doylesburg. The earlier use of Hatter and Snyder in south-central Pennsylvania was found to be a misapplication of the central Pennsylvania terms. The Fannettsburg and Doylesburg approximate the previous usage. The Mercersburg is divided into two new members, the Housum and Kauffman.

The Pinesburg and Fannettsburg members of the Shippensburg are Hatterian in age. The Pinesburg thins westward by overlap and convergence from Champlain miogeosynclinal sections in the eastern belts of outcrop to sections on the Adirondack line in Path Valley. From Path Valley, the Pinesburg equivalents thicken westward to Blacklog Valley in the Allegheny belt. Thick Fannettsburg sections in the eastern belts near Marion converge abruptly to thin shallow-water deposits in the western belts and maintain about the same thickness to Blacklog Valley. The Doylesburg member of Lowville (lower Hunterian) age thins irregularly eastward from Blacklog Valley and has not been recognized in the eastern belts. This thickness variation is interpreted as the result of erosion on the pre-Mercersburg unconformity. Beds of upper Hunterian age are absent throughout the area of study.

The Mercersburg formation of Rockland and Kirkfield age shows continued existence of all the structural features present during the deposition of the lower part of the Shippensburg. The Housum member has been identified only in the eastern

belts of outcrop in Pennsylvania, although a few feet of beds occupy the same stratigraphic position in some of the western belts. This westward thinning may have been accomplished by overlap or convergence or both. The Kauffman member thins westward to Path Valley, probably by convergence. The equivalents of both members thicken again toward central Pennsylvania.

The Oranda formation of Shoreham age is thickest to the northeast in Cumberland County, thereby contrasting with the older formations which have their greatest development in southeastern exposures. The Oranda thins westward to Path Valley at least partly by convergence. Equivalents of the upper beds diverge again towards central Pennsylvania, but there is no thickening of the lower beds. Thus, during the early part of Oranda time the structure along the Adirondack line was a monoclinal flexure with greater relative depression and deposition in the Champlain miogeosyncline. Later the structure resumed the character of an arch. The first appearance of argillaceous material in the Middle Ordovician is in the southeastern-most exposures of the Oranda in Franklin County, Pennsylvania. It is interpreted as the result of the Vermontian disturbance and uplift to the east. Convergence from the more westerly limestone beds to the eastern argillaceous beds suggests approach to marginal conditions.

The upper beds of the lower Martinsburg are poorly exposed in the eastern belts of outcrop. These beds maintain a rather uniform thickness through the western belts of outcrop in the Cumberland Valley, converge sharply in northern Path Valley, and diverge again in the Salona limestone of Blacklog Valley.

Relative to the Champlain miogeosyncline, a positive structural feature of variable prominence was along the Adirondack line throughout the deposition of most of the Shippensburg, all of the Mercersburg, Oranda, and Lower Martinsburg. It served as an arch, a belt of relatively less depression, throughout the deposition of all of these formations except the lower part of the Oranda and possibly the Doylesburg member of the Shippensburg. East of the Adirondack line, a monoclinal flexure, produced by greater depression to the east, permitted the accumulation of the relatively thick Fannettsburg, Mercersburg, and Oranda beds in the eastern belts.

GEOLOGIC SECTIONS

The following stratigraphic sections are arranged by belts of outcrop from northwest to southeast. Within each belt the sections are arranged from north to south. The figures in column I are the thicknesses of the individual described units. The figures in column II are the cumulative thicknesses in the member or formation as the case may be.

Lower Middle Ordovician sections in Blacklog Valley, Huntingdon County, Pennsylvania

1. "North" Quarry section: Shallow quarry on Gilland farm, 1.8 mi. S. 63°E. of Orbisonia, Pa. and 0.3 mi. N. 20°E. of the Grove (Binkley and Ober) Quarry. Beds dip 40°, N. 65°W.

	Thickness (Feet) I	II
Salona limestone:		
Metabentonite, yellow; with plates of secondary calcite	0.7	37.2
Limestone, black to dark brown, fine- to medium-grained, argillaceous, platy to slabby, contains *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens)	11.7	36.5

	Thickness (Feet) I	II
Metabentonite and shale unit; Salona 2 (0.3′ brown shale; 0.7′ yellow metabentonite)	1.0	23.8
Calcilutite, dark gray to black, platy to slabby, *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens)	15.5	22.8
Limestone, dark gray, fine- to medium-grained, irregular platy bedded, blue weathering	2.5	7.3
Limestone, medium gray, medium-grained to medium crystalline, gray granular weathering, *Echinosphaerites*	4.0	4.8
Metabentonite, yellow; Salona 0	0.3	1.8
Limestone, medium gray, medium-grained to medium crystalline, gray granular weathering	1.5	1.5
Nealmont formation:		
Rodman-Centre Hall members (undivided):		
Limestone, medium to dark gray, platy, crinkly bedded, blue weathering, *Leptaena* sp. cf. *L. charlottae* Winchell and Schuchert	52.2	52.5
Metabentonite, yellow, N_2?	0.3	0.3
Oak Hall member:		
Limestone, medium to dark gray, fine-grained crinkly to slabby, blue and rubbly weathering, contains *Lambeophyllum*. Black chert nodules in upper 2 feet	25.1+	25.1+
Lower part of member poorly exposed	—	—

2. Grove (Binkley and Ober) Quarry section: 1.8 mi. S. 54°E. of Orbisonia, Pa. Section measured to top of exposure in northern re-entrant of quarry. Beds dip 45°, N. 55°W.

	Thickness (Feet) I	II
Salona limestone:		
Calcilutite, black, platy to slabby	5.5+	12.2+
Limestone, dark gray to brown, medium-grained to medium crystalline, platy to slabby, gray granular weathering	4.3	6.7
Metabentonite, yellow and white, shaly; Salona O	0.8	2.4
Limestone, dark gray to brown, medium-grained to medium crystalline, platy to slabby, gray granular weathering	1.6	1.6
Nealmont limestone:		
Rodman and Centre Hall members (undivided):		
Limestone, dark gray, fine- to medium-grained, thin, crinkly to slabby bedded, bluish weathering	58.0	58.5
Metabentonite, yellow and white; N_2 (?)	0.5	0.5
Oak Hall Member:		
Limestone, dark gray, fine- to medium-grained, crinkly bedded, blue weathering; lower 5 feet quite rubbly and fossiliferous, upper 1 foot contains small black chert nodules	26.2	26.2
Benner limestone:		
Snyder member:		
Calcilutite, dove to light gray, massively bedded, white weathering, *Tetradium* sp.	19.8	50.4
Limestone, dark to medium gray, fine-grained, crinkly bedded, bluish weathering; contains *Tetradium cellulosum* (Hall) and *T. fibratum* Safford	11.8	30.6
Calcilutite, dove to light gray, massively bedded, white weathering	4.5	18.8
Limestone, thin platy to shaly	0.1	14.3
Limestone, light gray to dove, medium-grained to finely crystalline, faintly laminated, slabby evenly bedded, light-gray weathering	6.7	14.2
Limestone, medium gray, medium-grained to finely crystalline, faintly laminated, slabby evenly bedded, gray to buff weathering, contains some limestone pebbles in lower 3 feet	7.5	7.5
Hatter limestone:		
Hostler member:		
Limestone, medium gray, medium-grained to finely crystalline, slabby evenly bedded, light gray to buff weathering	6.5	47.1

	Thickness (Feet) I	II
Limestone, dark gray to black, medium-grained, irregular slabby; contains *Girvanella*; upper 1 foot finely laminated	18.4	40.6
Limestone, dark gray, fine- to medium-grained, crinkly bedded, blue weathering	7.5	22.2
Grazier member:		
Limestone, dark to medium gray, fine- to medium-grained, crinkly bedded, blue weathering	6.8	14.7
Limestone, dark gray, medium-grained, prominent irregular silty partings, dark brown cobbly to rubbly weathering	7.9	7.9
Loysburg formation:		
Clover member:		
Limestone, dark to medium gray, fine- to medium-grained, stylolitic, crinkly bedded, light-gray weathering	12.9	12.9

3. St. Mary's Church section: 2.5 miles S. 20°E. of Orbisonia, Pa. Measured along farm lane extending southwest from U. S. Route 522.

	Thickness (Feet) I	II
Salona limestone:		
Covered	—	—
Calcilutite, black	0.5	99.8
Metabentonite, yellow and white; Salona 6	0.3	99.3
Calcilutite, black, slabby to massively bedded, in part highly cleaved and shaly weathering	48.4	99.0
Metabentonite unit; Salona 5 (0.2 black clay; 0.3 white metabentonite; 0.6 yellow shaly metabentonite; 0.5 white metabentonite)	1.6	50.6
Calcilutite, black, slabby to massively bedded, in part highly cleaved and shaly weathering; black silicified surface at top; contains *Cryptolithus*	26.1	49.0
Metabentonite and shale unit; Salona 4 (0.3 black shale; 0.6 yellow metabentonite)	0.9	22.9
Shale, dark gray to brown, calcareous, highly cleaved, with a few heavy beds of dark-gray to black calcilutite; contains *Sinuites*	20.5	22.0
Metabentonite unit; Salona 3 (0.7 lemon yellow metabentonite; 0.8 yellow brown clay)	1.5	1.5
Lower part of formation covered	—	—
Nealmont limestone:		
Rodman and Centre Hall members (undivided):		
Upper part of formation covered. Total thickness between exposures	69.0	—
Limestone, dark gray, fine- to medium-grained, crinkly to slabby, blue weathering; interval partly covered	43.1	43.3
Metabentonite, yellow clay, poorly exposed; N_2	0.2	0.2
Oak Hall member:		
Limestone, dark gray, fine- to medium-grained, crinkly to slabby, blue weathering; interval partly covered, remnants of highly weathered black silicified surface at top	20.7	20.7
Lower part of formation covered	—	—
Benner limestone:		
Snyder member:		
Covered interval	—	—
Hatter limestone:		
Upper part of formation covered. Total thickness between exposures	97.0	—
Limestone, dark gray, argillaceous, brown rubbly weathering	1.5	4.0
Limestone, dark to medium gray, cobbly, blue weathering	2.5	2.5
Loysburg formation:		
Clover member:		
Limestone, medium gray, fine-grained, crinkly bedded, white weathering	11.0	

Lower Middle Ordovician sections in Path Valley, Franklin County, Pennsylvania

4. **Doylestown Quarry section**: 1.5 mi. S. 47°W. of Doylesburg and 0.3 mi. W. of Pennsylvania Route 75. Composite section compiled from measurements at both southern and northern quarry openings. Beds dip 35°, N. 40°W.

	Thickness (Feet) I	II
Martinsburg shale:		
Shale, black, platy, calcareous, graptolitic; contains *Brongniartella*	2.5	21.4
Metabentonite, yellow; Salona 4	1.2	18.9
Limestone, black, fine-grained, platy to slabby, argillaceous; top bedding surface silicified	2.7	17.7
Limestone, black, fine-grained, highly cleaved, argillaceous, brownish weathering	3.5	15.0
Limestone, black, fine-grained, highly cleaved, argillaceous, brown weathering; contains *Sinuites*	1.4	11.5
Metabentonite and shale unit; Salona 3 (0.1 brown shale; 0.2 yellow metabentonite; 0.3 greenish-yellow metabentonite)	0.6	10.1
Limestone, black, medium grained, crinoidal; weathers to purplish cast; contains *Sinuites*, *Lingula*	5.0	9.5
Metabentonite and shale unit; Salona 2 (0.2 yellow metabentonite; 0.5 brownish shale; 0.7 yellow shaly metabentonite)	1.4	4.5
Limestone, black, medium-grained, crinoidal, argillaceous; weathers to purplish cast	2.6	3.1
Conglomerate; dark gray, fine-grained limestone pebbles up to 3 inches in diameter in a dark gray to black, semi-crystalline limestone matrix	0.5	0.5
Oranda formation:		
Limestone, black, fine-grained, semicrystalline, cobbly, argillaceous, with *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens)	1.9	26.3
Calcilutite, black, cobbly; with a few medium-gray, medium-crystalline interbeds, contains *Lingula*	5.9	24.4
Metabentonite, white and yellow, Salona 0	0.7	18.5
Calcilutite, dark gray to black, slabby to massive; with thin silty partings; *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens), *Receptaculites* sp.	17.8	17.8
Lower part of formation covered	—	—
Mercersburg formation:		
Upper part of formation covered. Total thickness between exposures	12.0	—
Limestone, dark to medium gray, fine- to medium-grained, crinkly bedded; becomes slabby upwards, bluish weathering; contains *Sowerbyella* sp. of *S. punctistriata* (Mather)	36.2	50.9
Limestone, dark to medium gray, fine- to medium-grained; with crinkly silty partings, bluish and rubbly weathering	14.7	14.7
Shippensburg formation:		
Doylesburg member:		
Calcilutite, light to dark dove, massively bedded, white weathering; partially concealed; contains *Tetradium cellulosum* (Hall)	38.5	38.5

5. **Dry Run Quarry section**: measured at Binkley and Ober Quarry in southeast environs of Dry Run, Pa. Beds dip 30°, N. 50°W.

	Thickness (Feet) I	II
Oranda formation:		
Limestone, black, fine-grained, slabby	1.0	21.4
Metabentonite, light yellow; Salona 0	0.3	20.4
Limestone, black, fine-grained, slabby; bottom foot semicrystalline; lower contact not well exposed	20.1	20.1
Mercersburg formation:		
Limestone, dark gray, fine- to medium-grained, crinkly to rubbly bedded, blue weathering; black chert nodules in lower 2 feet	42.9	50.4

	Thickness (Feet) I	II
Metabentonite unit; Mercersburg 2 (0.7 lemon-yellow metabentonite; 0.3 yellow shaly metabentonite)	1.0	7.5
Limestone, dark to medium gray, rubbly, blue weathering; black chert nodules in top foot	2.5	6.5
Limestone, dark to medium gray, medium-grained, semicrystalline; with stylolitic partings	4.0	4.0
Shippensburg formation:		
Doylesburg member:		
Calcilutite, dove to dark gray, massive, white weathering; contains *Tetradium cellulosum* (Hall) and *Cryptophragmus antiquatus* Raymond	39.8	39.8
Fannettsburg member:		
Limestone, dark to medium gray, fine-grained, rubbly to platy bedded; with thick, medium-gray, medium-grained limestone partings	8.2	23.3
Limestone, dark to medium gray, medium-grained to finely crystalline; contains *Solenopora* sp. and *Stromatocerium* sp.	15.1	15.1
Pinesburg member:		
Limestone, dark to medium gray, fine- to medium-grained; in crinkly to cobbly beds; contains *Girvanella* sp.	11.5	29.3
Limestone, medium gray, medium-grained to finely crystalline, cross-laminated	1.8	17.8
Limestone, dark gray, fine-grained, thin irregularly bedded; with thick, dark-gray, medium-grained limestone partings	16.0	16.0
"Stones River" limestone:		
Calcilutite, dove to medium gray, massively bedded, white weathering; contains *Tetradium syringoporoides* Ulrich	7.8	23.7
Calcilutite, dove; with light-gray, buff-weathering laminae; contains *Tetradium syringoporoides* Ulrich	15.9	15.9

6. Spring Run Quarry section: measured at Hammond Quarry, 0.8 miles S. 5°E. of Spring Run.

	Thickness (Feet) I	II
Martinsburg shale:		
Shale, black, calcareous, highly cleaved, buff to brown weathering	7.0	13.3
Metabentonite, light-yellow clay; Salona 3	0.1	6.3
Limestone, black, argillaceous, crinoidal; contains *Sinuites*	3.9	6.2
Metabentonite and shale unit; Salona 2 (0.1 yellow metabentonite; 0.2 brown shale, white calcite plates abundant throughout)	0.3	2.3
Limestone, dark gray to black, fine- to medium-grained, cobbly	1.0	2.0
Calcarenite, medium to dark gray, medium- to coarse-grained; with a conglomerate bed containing dark-gray, fine-grained limestone pebbles at the top	1.0	1.0
Oranda formation:		
Limestone, dark gray to black, fine-grained, argillaceous, and crinoidal	2.1	17.9
Metabentonite and shale unit; Salona O (0.1 brown shale; 0.1 white and light-yellow metabentonite with white calcite plates)	0.2	15.8
Limestone, dark gray to black, fine-grained, argillaceous and crinoidal; black chert nodules in upper 2 feet; contains *Receptaculites* sp.	13.6	15.6
Limestone, dark gray to black, medium-grained to medium crystalline, gray, granular weathering; contains *Receptaculites* sp. and *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens). Bottom contact distinct, annealed, fills shallow depression in underlying bed	2.0	2.0
Mercersburg formation:		
Limestone, dark to medium gray, fine-grained, irregular platy with crinkly partings; quite fossiliferous	41.3	47.2
Metabentonite unit; Mercersburg 2 (0.1 light-yellow metabentonite; 0.7 gray, shaly metabentonite; 0.2 yellow metabentonite with white calcite plates)	1.0	5.9

	Thickness (Feet) I	II
Limestone, dark gray, fine-grained, massive; with faint crinkly partings, blue weathering; top bedding surface is black and silicified	4.3	4.9
Calcarenite, medium gray, coarse; on channelled disconformable surface	0.6±	0.6±
Shippensburg formation:		
Doylesburg member:		
Limestone, medium gray, fine- to medium-grained, light-gray weathering; contains *Cryptophragmus antiquatus* Raymond	3.1	22.5
Calcilutite, medium gray to dark dove	5.8	19.4
Calcilutite, light dove, massively bedded, white weathering; contains *Tetradium cellulosum* (Hall)	13.6	13.6
Fannettsburg member:		
Limestone, light to medium gray, fine- to medium-grained, laminated, light-gray weathering; contains *Tetradium cellulosum* (Hall)	2.3	28.3
Limestone, dark to medium gray, fine- to medium-grained, laminated, medium-gray weathering; with thin silty partings; contains *Tetradium cellulosum* (Hall), *Solenopora*	2.8	26.0
Limestone, dark gray, argillaceous, with *Solenopora*	4.0	23.2
Limestone, dark gray, fine- to medium-grained, crinkly bedded, blue weathering; with *Solenopora* in abundance	4.2	19.2
Limestone, dark gray, medium-grained to finely crystalline; in platy beds with anastomosing argillaceous partings; weathers gray and granular with a ribbony appearance	4.4	15.0
Limestone, dark to medium gray, fine-grained, slabby bedded; contains *Girvanella* and *Solenopora*	8.5	10.6
Calcarenite, medium gray, medium to coarsely crystalline, slabby	2.1	2.1
Pinesburg member:		
Limestone, dark to medium gray, fine-grained, slabby bedded; contains *Girvanella*	6.1	21.1
Limestone, dark to medium gray, medium-grained, slabby; with prominent argillaceous limestone partings	10.5	15.0
Limestone, medium gray, coarse-grained to medium crystalline, slabby	4.5	4.5
"Stones River" limestone:		
Calcilutite, medium to dark gray, platy to massive; with *Tetradium syringoporoides* Ulrich	1.5+	1.5+

7. **Willow Hill-Turnpike section:** Compiled from measurements on both sides of the Pennsylvania Turnpike, just east of Pennsylvania Route 75. Beds dip 50°, S. 60°E.

	Thickness (Feet) I	II
Martinsburg shale:		
Limestone, black, argillaceous, buff shaly weathering; with a few thin beds of black calcilutite; contains graptolites	—	—
Metabentonite, yellow, with white calcite plates; Salona 7	0.2	58.8
Limestone, black, argillaceous, buff shaly weathering; contains *Cryptolithus*, *Corynoides*	7.1	58.6
Metabentonite and shale unit; Salona 6 (0.1 brown shale; 0.5 yellow metabentonite)	0.6	51.5
Limestone, black, argillaceous, buff shaly weathering; with a few beds of black calcilutite; contains *Brongniartella*, *Cryptolithus*	30.5	50.9
Metabentonite and shale unit; Salona 5 (0.2 yellow-brown shale; 0.2 light-yellow metabentonite)	0.4	20.4
Limestone, black, argillaceous, buff shaly weathering; with a few beds of black calcilutite; contains *Brongniartella*	5.4	20.0
Metabentonite and shale unit; Salona 4 (0.1 yellow papery metabentonite; 0.1 brown shale; 0.1 yellow papery metabentonite)	0.3	14.6
Limestone, black, argillaceous, brown to buff shaly weathering; top surface black and silicified	8.7	14.3

	Thickness (Feet) I	II
Metabentonite and shale unit; Salona 3 (0.1 brown shale; 0.1 yellow metabentonite; 0.3 greenish-yellow metabentonite)	0.5	5.6
Limestone, black, argillaceous; weathers brown to buff shaly rubble	2.6	5.1
Limestone, dark gray to black, argillaceous, semicrystalline	0.6	2.5
Metabentonite and shale unit; Salona 2 (0.5 yellow and white metabentonite; 0.1 brown shale; 0.1 white metabentonite)	0.7	1.9
Limestone, black, fine-grained, crinoidal, buff weathering	1.2	1.2
Oranda formation:		
Calcilutite, black; with interbeds of medium gray, medium-grained to coarsely crystalline limestone	4.0	18.1
Metabentonite and shale unit; Salona 0 (0.1 brown shale; 0.2 yellow metabentonite)	0.3	14.1
Limestone, dark gray to black, argillaceous, crinoidal, cobbly bedded	0.8	13.8
Calcilutite, dark gray to black, platy to slabby, buff to blue weathering; contains *Receptaculites* sp.; lower beds partly covered	13.0	13.0
Mercersburg formation:		
Limestone, medium gray, fine-grained, thin crinkly to slabby, blue weathering; with prominent argillaceous partings; lower beds partly covered	33.2	33.2
Shippensburg formation:		
Doylesburg member:		
Calcilutite, dove, massively bedded, white weathering; with *Tetradium cellulosum* (Hall); interval partly covered	22.2	22.2
Fannettsburg member:		
Limestone, light gray, medium-grained, faintly laminated, light-gray weathering	5.0	29.6±
Limestone, dark gray, medium-grained, mottled	3.1	24.6±
Limestone, dark gray, medium-grained, argillaceous; contains *Solenopora* sp.; interval partly covered	21.5±	21.5±
Pinesburg member:		
Limestone, medium to dark gray, fine- to medium-grained, thin bedded; interval partly covered	12.7±	12.7±
"Stones River" limestone:		
Calcilutite, dove, massive, white weathering; contains *Tetradium syringoporoides* Ulrich	5.5	5.5

8. Metal section: Section measured in farm lane just east of Pennsylvania route 75 and 1.8 miles N. 58°E. of Metal. Beds dip 40°, S. 55°E.

	Thickness (Feet) I	II
Martinsburg shale:		
Limestone, black, fine-grained, argillaceous, highly fractured, buff shaly weathering	26.5	192.3
Clay, yellow brown, probably metabentonitic	0.2	165.8
Limestone, black, fine-grained, argillaceous, highly fractured, buff shaly weathering	12.8	165.6
Metabentonite and shale unit; (0.1 brown clay shale; 0.2 yellow gritty metabentonite)	0.3	152.8
Limestone, black, fine-grained, argillaceous, gray to buff, splintery weathering; with slaty cleavage	16.8	152.5
Metabentonite, cream-colored; with white calcite plates; Salona 7 (?)	0.2	135.7
Limestone, black, fine-grained, argillaceous, buff weathering; with slaty cleavage	6.4	135.5
Metabentonite and shale unit; Salona 6 (?) (0.1 brown shale; 0.2 yellow and white metabentonite)	0.3	129.1
Limestone, black, fine-grained, argillaceous, buff weathering; with slaty cleavage; lower 57 feet partly covered	65.9	128.8
Metabentonite, white and light yellow; Salona 5 (?)	1.0	62.9

	Thickness (Feet) I	II
Limestone, black, fine-grained, buff weathering; with slaty cleavage; interval partly covered	11.8	61.9
Metabentonite, yellow; Salona 4 (?)	0.1	50.1
Limestone, black, fine-grained, buff weathering; with slaty cleavage	3.7	50.0
Covered interval	43.6	46.3
Limestone, black, fine-grained, argillaceous, slabby	0.5	2.7
Metabentonite unit; Salona 2 (0.5 light- to deep-yellow metabentonite; 0.5 light-gray to cream metabentonite)	1.0	2.2
Limestone, dark gray, fine-grained, crinoidal	1.2	1.2
Oranda formation:		
Limestone, black, fine-grained, slabby, gray weathering; interval partly covered	17.1	17.1
Mercersburg formation:		
Limestone, dark gray, fine-grained, crinkly bedded, blue and rubbly weathering	39.9	39.9
Shippensburg formation:		
Doylesburg member:		
Limestone, dark gray to dark dove, fine- to medium-grained, slabby bedded, gray weathering	2.9	14.9
Calcilutite, dove, massive, white weathering; contains *Tetradium cellulosum* (Hall)	10.4	12.0
Limestone, dove, very fine-grained, whitish weathering	1.6	1.6
Fannettsburg member:		
Limestone, light gray to dark dove, coarse-grained to medium crystalline, slabby, light-gray weathering	3.3	29.1
Limestone, dark gray, medium crystalline; with interbeds of medium-gray, coarsely crystalline limestone; platy to slabby, gray weathering	8.6	25.8
Limestone, dark gray, fine-grained; with argillaceous partings, and interbeds of medium-gray, coarse calcarenite	4.7	17.2
Limestone, medium gray, medium to coarsely crystalline, platy to slabby, gray weathering	12.5	12.5
Pinesburg member:		
Limestone, dark gray, fine-grained, cobbly to crinkly, blue weathering; containing a few beds of *Girvanella* conglomerate	11.9	16.5
Limestone, dark gray to black, cobbly; with argillaceous partings	4.6	4.6
Bottom of member unexposed	—	—

Lower Middle Ordovician sections in the Fort Loudon—Blue Spring belt of outcrop, Franklin County, Pennsylvania

9. Fort Loudon Quarry Section: Small quarry 0.4 miles south of Fort Loudon and west of Pennsylvania Route 75.

	Thickness (Feet) I	II
Shippensburg formation:		
Doylesburg member:		
Upper part of member covered	—	—
Calcilutite, dove, massive, white weathering; contains abundant *Tetradium cellulosum* (Hall)	1.1	2.6
Conglomerate, small pellets of dove calcilutite; contains abundant *Tetradium cellulosum* (Hall)	1.5	1.5
Fannettsburg member:		
Limestone, dove to light gray, fine- to medium-grained, cross-laminated, slabby, light-gray granular weathering	5.0	23.6
Limestone, medium to dark gray, fine- to medium-grained, cross-laminated, slabby to platy, medium gray granular weathering	4.3	18.6
Shale, calcareous; with thin lenses of medium-gray, medium-grained, platy, fossiliferous limestone; abundant *Cryptophragmus antiquatus* Raymond	0.3	14.3
Limestone, dark to medium gray, medium-grained, cross-laminated, slabby; with a few interbeds of medium-gray calcarenite; contains *Caryocystites, Glyptorthis, Favistella*	2.4	14.0

	Thickness (Feet) I	II
Calcarenite, medium gray, coarse, slabby, medium-gray weathering; with a few interbeds of dark to medium-gray, medium-grained limestone; *Solenopora* abundant	5.7	11.6
Limestone, dark gray, fine- to medium-grained, slabby, gray weathering; with a few interbeds of medium-gray, coarse calcarenite	5.9	5.9
Pinesburg member:		
Limestone, dark gray, fine- to medium-grained, blue-gray weathering; with irregular argillaceous seams; *Girvanella* abundant in top foot	5.4	5.9
Limestone, light gray to tan, fine- to medium-grained, buff weathering	0.5	0.5
Lower part of member unexposed	—	—

10. Fort Loudon field section: Compiled from scattered exposures on the south side of a small valley, 0.6 miles south of Fort Loudon between Pennsylvania Route 75 and Cumberland Valley Railroad. Beds dip 15°, N. 55°E.

	Thickness (Feet) I	II
Mercersburg formation:		
Upper part poorly exposed	—	—
Limestone, dark gray, fine-grained to medium crystalline	2.5	2.5
Shippensburg formation:		
Doylesburg member:		
Calcilutite, light to dark dove; slabby in upper part, with dark- to medium-gray, fine-grained, crinkly limestone in lower part	4.2	15.8
Calcilutite, light to dark dove, platy	0.9	11.6
Conglomerate; small pellets of dark dove calcilutite; massively bedded; contains *Cryptophragmus*	1.4	10.7
Limestone, medium gray, fine- to medium-grained, platy; abundant *Cryptophragmus* in bed near top	2.6	9.3
Calcilutite, dove, massive; with a bed of dove, calcilutite pellet conglomerate at top	1.4	6.7
Covered interval	3.4	5.3
Calcilutite, dove, massively bedded; with abundant *Tetradium cellulosum* (Hall)	1.9	1.9
Base of member covered	—	—
Fannettsburg member:		
Upper part of member covered. Total thickness between exposures	16.7	—
Limestone, light to medium gray, laminated, platy; with interbeds of coarse calcarenite containing *Solenopora*	3.9	3.9
Base of member covered	—	—
Pinesburg member:		
Upper part of member covered. Total thickness between exposures	2.0	—
Limestone, dark gray, cobbly to platy, blue rubbly weathering	13.5	40.8
Limestone, dark gray, cobbly, blue to buff rubbly weathering; with prominent argillaceous partings	27.3	27.3
Lower part of member covered	—	—
"Stones River" limestone:		
Upper part of formation covered. Total thickness between exposures	21.3	—
Calcilutite, dove; and dove, medium-grained limestone, platy to massively bedded, white weathering	5.9	5.9
Underlying beds poorly exposed	—	—

11. Africa section: Compiled from measurements in road cuts 1.4 miles S. 82°W. of Mercersburg and 1.6 miles S. 63°W. of Mercersburg. Beds dip 40°, S. 85°E.

	Thickness (Feet) I	II
Martinsburg shale:		
Limestone, black, fine-grained, argillaceous, buff shaly weathering	17.3	219.6
Metabentonite, yellow	0.2	202.3
Limestone, black, fine-grained, argillaceous, buff shaly weathering	3.2	202.1

	Thickness (Feet) I	II
Metabentonite and shale unit; (0.6 brown calcareous shale; 0.2 yellow and gray metabentonite)	0.8	198.9
Limestone, black, fine-grained, argillaceous, buff shaly weathering	5.7	198.1
Metabentonite unit; (0.2 light-gray metabentonite; 0.3 yellow metabentonite; 0.3 light-yellow metabentonite)	0.8	192.4
Limestone, black, fine-grained, argillaceous, buff shaly weathering	18.2	191.6
Metabentonite, yellow, gritty	0.2	173.4
Limestone, black, fine-grained, argillaceous, buff shaly weathering	9.4	173.2
Clay, yellow brown to light yellow; probably metabentonite	0.1	163.8
Limestone, black, fine-grained, argillaceous, shaly; this interval and succeeding beds are partly covered	6.1	163.7
Metabentonite unit; Salona 7 (?) (0.2 light-yellow metabentonite; 0.3 deep-yellow metabentonitic shale; a little white metabentonite; 0.3 light-yellow metabentonite)	0.8	157.6
Limestone, black, fine-grained, argillaceous, buff shaly weathering	6.8	156.8
Metabentonite and shale unit; Salona 6 (?) (0.1 light-yellow metabentonite; 0.2 yellow shale; 0.3 light-yellow metabentonite)	0.6	150.0
Limestone, black, fine-grained, argillaceous, buff shaly weathering	17.6	149.4
Metabentonite unit (0.1 light-yellow metabentonite; 0.1 white calcite plates)	0.2	131.8
Limestone, black, fine-grained, argillaceous, buff shaly weathering	9.7	131.6
Metabentonite, light yellow, with white calcite plates	0.2	121.9
Limestone, black, fine-grained, argillaceous, buff shaly weathering	6.1	121.7
Metabentonite and shale unit; (0.1 light-yellow metabentonite with white calcite plates; 0.1 gray-brown shale; 0.1 deep-yellow shaly metabentonite)	0.3	115.6
Limestone, black, fine-grained, argillaceous, buff shaly weathering	1.4	115.3
Metabentonite, light yellow; with white calcite plates	0.1	113.9
Limestone, black, fine-grained, argillaceous, buff shaly weathering	4.9	113.8
Metabentonite and shale unit; (0.1 brown clay shale; 0.3 yellow metabentonite; 0.4 light-yellow metabentonite with white calcite plates)	0.8	108.9
Limestone, black, fine-grained, argillaceous, buff shaly weathering; top bedding surface is black and silicified	16.7	108.1
Metabentonite, medium yellow; with white calcite plates	0.1	91.4
Shale, dark gray, calcareous, fissile	0.8	91.3
Limestone, black, fine-grained, argillaceous, massively bedded, buff weathering	13.0	90.5
Metabentonite and shale unit; Salona 5 (?) (0.1 deep-yellow metabentonite and gray brown shale; 0.2 yellow metabentonite; 0.2 light-yellow metabentonite)	0.5	77.5
Limestone, black, fine-grained, argillaceous, massively bedded, buff weathering	6.6	77.0
Metabentonite unit; Salona 4 (?) (0.3 white and deep-yellow metabentonite; 0.2 yellow-green metabentonite with white calcite plates)	0.5	70.4
Limestone, black, fine-grained, argillaceous, massive, buff weathering	43.4	69.9
Metabentonite unit; Salona 3 (0.1 gray-green shaly metabentonite; 0.1 deep-yellow shaly metabentonite; 0.2 gray-green metabentonite)	0.4	26.5
Limestone, black, fine-grained, argillaceous, massive, buff weathering	11.7	26.1
Metabentonite, yellow and white; Salona 2	0.4	14.4
Limestone, black, fine-grained, argillaceous, shaly weathering; contains *Cryptolithus*	1.9	14.0
Metabentonite, white; Salona 1	0.1	12.1
Limestone, black, fine-grained, argillaceous, slabby; contains *Cryptolithus*; top bedding surface black and silicified	10.9	12.0
Limestone, dark gray, semicrystalline, gray platy weathering; interval poorly exposed	1.1	1.1
Oranda formation:		
Limestone, black, medium-grained, slabby; with shaly partings; weathers to a purplish cast; contains *Lingula*, *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens)	6.0	21.6

	Thickness (Feet) I	II
Metabentonite and siltstone unit; Salona 0 (0.2 olive-drab blocky siltstone; 0.5 yellow and white metabentonite)	0.7	15.6
Calcilutite, dark gray to black, slabby; with even shaly partings	4.9	14.9
Metabentonite and shale unit (0.1 brown shale; 0.2 light-yellow metabentonite with white calcite plates)	0.3	10.0
Calcilutite, dark gray to black, slabby; with even shaly partings	6.8	9.7
Limestone, dark gray, medium-grained to medium crystalline, platy to slabby, dark-gray weathering; contains *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens)	2.9	2.9
Mercersburg formation:		
Limestone, dark gray, fine-grained, platy to slabby	25.3	62.0
Limestone, dark gray, fine-grained, crinkly bedded, blue weathering	16.2	36.7
Metabentonite and siltstone units; Mercersburg 2 (1.0 yellow-brown clay; 2.0 olive-drab blocky siltstone; 1.3 yellow and white metabentonite; 0.3 olive-drab siltstone; 0.4 yellow and white metabentonite)	5.0	20.5
Limestone, dark gray, fine- to medium-grained, crinkly bedded, blue weathering	15.5	15.5
Shippensburg formation:		
Doylesburg member:		
Limestone, dark dove, fine- to medium-grained, massive; contains *Tetradium* sp.	8.0	9.2
Calcilutite, dove, massively bedded, white weathering	1.2	1.2
Lower part of member covered	—	—
Fannettsburg member:		
Upper part of member covered. Total thickness between exposures.	10.1	—
Limestone, medium gray, medium-grained to finely crystalline, slabby	4.7	6.2
Calcarenite, dark gray, coarse, crinoidal, massive	1.5	1.5
Lower part of member covered	—	—
Pinesburg member:		
Upper part of member covered. Total thickness between exposures.	28.4	—
Limestone, dark gray, medium-grained, cobbly to slabby, blue weathering, highly fractured; may be faulted	7.1	7.1
"Stones River" limestone:		
Limestone, light to dark dove, fine- to medium-grained, massively bedded, white to buff weathering	6.3	6.3

12. Blue Spring Section: Compiled from measurements 2.8 miles S. 45°W. of Mercersburg. Beds dip 20°, S. 65°E.

	Thickness (Feet) I	II
Mercersburg formation:		
Upper part of formation covered	—	—
Limestone, dark gray, fine- to medium-grained, crinkly to platy	3.2	6.6
Calcarenite, dark gray, medium to coarse	3.4	3.4
Shippensburg formation:		
Doylesburg member:		
Limestone, dark gray, fine- to medium-grained	2.0	24.0
Limestone, dark dove to medium gray, fine- to medium-grained, slabby; *Cryptophragmus* abundant in bottom 4 inches	3.4	22.0
Limestone, medium gray to dark dove, fine- to medium-grained, platy	2.3	18.6
Calcilutite, dark to light dove, mottled, platy	3.6	16.3
Limestone, medium gray to dark dove, fine- to medium-grained, platy	2.0	12.7
Conglomerate, dark dove to tan, calcilutite-pellet; with a few interbeds of dark-dove to medium-gray, fine- to medium-grained limestone	7.7	10.7
Limestone, dark dove to medium gray, fine- to medium-grained, crinkly bedded; contains *Cryptophragmus* at bottom	3.0	3.0

	Thickness (Feet) I	II
Fannettsburg member:		
Limestone, medium gray, medium-grained, cross-laminated, gray weathering; with a few interbeds of coarse calcarenite; contains *Girvanella, Glyptorthis*	11.3	34.7
Covered interval	9.5	23.4
Calcarenite, medium gray, coarse	2.2	13.9
Lower part of member covered. Location of contact based on abundant float	11.7	11.7
Pinesburg member:		
Upper part of member covered	51.7	59.7
Limestone, dark to medium gray, fine- to medium-grained, cobbly to platy, blue weathering	8.0	8.0
Base of member covered	—	—

Lower Middle Ordovician sections in the Dickey belt of outcrop, Franklin County, Pennsylvania

13. Dickey Section: Compiled from exposures along the Cumberland Valley Railroad and northwest of the Fry Coal and Stone Co. Quarry, 0.6 miles N. 30°W. of Dickey. Beds dip 40°, N. 80°W.

	Thickness (Feet) I	II
Oranda formation:		
Calcarenite, light gray, coarse, massively bedded; contains *Echinosphaerites*	6.6	6.6
Mercersburg formation:		
Limestone, dark gray, fine-grained, crinkly to slabby, blue weathering	58.1	104.9
Covered interval	4.0	46.8
Limestone, dark gray, fine- to medium-grained, crinkly to slabby, blue weathering; partly covered	18.3	42.8
Covered interval	15.4	24.5
Metabentonite and siltstone unit; Mercersburg 2 (1.9 olive-drab siltstone; 0.9 yellow shaly metabentonite)	2.8	9.1
Limestone, dark gray, fine- to medium-grained, crinkly to cobbly, blue weathering; interval partly covered	6.3	6.3
Shippensburg formation:		
Doylesburg member:		
Calcilutite, dark dove, massively bedded; contains abundant *Tetradium cellulosum* (Hall)	6.3	28.6
Calcilutite, dark to light dove, massively bedded; contains abundant *Cryptophragmus* near top	10.4	22.3
Limestone, dark gray, fine- to medium-grained, crinkly bedded; some fine limestone pellet conglomerate in upper part	3.9	11.9
Covered interval	4.1	8.0
Limestone, dark dove, fine-grained; with *Tetradium cellulosum* (Hall); and 1′ dark-dove fine calcilutite-pellet conglomerate near bottom	3.9	3.9
Fannettsburg member:		
Calcarenite, medium to dark gray, medium to coarse, gray granular weathering, slabby	8.0	8.0
Lower part of member covered	—	—
Pinesburg member:		
Upper part of member covered. Total thickness between exposures	31.0	—
Limestone, dark gray-, fine- to medium-grained; with prominent buff weathering silty seams; abundant float only, no exposures	48.0	90.9
Limestone, dark gray, fine- to medium-grained, cobbly bedded, blue weathering	30.0	42.9
Covered interval	9.6	12.6
Calcarenite, medium gray, coarse, massively bedded; underlying contact is disconformable with 6 inches of relief	3.0	3.0
"Stones River" limestone:		
Limestone, dark gray, fine-grained, crinkly bedded; and dove, massively bedded calcilutite	—	—

Lower Middle Ordovician section in the Claylick-Church Hill belt of outcrop, Franklin County, Pennsylvania

14. Shimpstown Section: Exposure along the West Branch of Conococheague Creek, 1.5 miles N. 65°E. of Shimpstown and 2.5 miles S. 38°E. of Mercersburg. Beds dip 75°, N.80°W.

	Thickness (Feet) I	II
Martinsburg shale:		
Limestone, black, fine-grained, platy to slabby, argillaceous, buff weathering; poorly exposed	—	—
Metabentonite, cream to light yellow; poorly exposed and probably slumped to give exaggerated thickness; Salona 2	3.2	8.3
Limestone, black, fine-grained to semicrystalline; *Sinuites* at top	2.8	5.1
Calcarenite, dark to medium gray, medium to coarse, slabby	2.3	2.3
Oranda formation:		
Calcilutite, black, slabby; contains *Receptaculites* near top	12.7	14.8
Limestone, medium to dark gray, medium-grained, semicrystalline, slabby, dark-gray weathering	2.1	2.1
Mercersburg formation:		
Limestone, medium to dark gray, fine- to medium-grained, crinkly bedded, blue weathering	45.1	53.4
Limestone, dark to medium gray, medium grained, semi- to medium crystalline, slabby	8.3	8.3
Shippensburg formation:		
Doylesburg member:		
Calcilutite, light dove, massively bedded, white weathering	22.5	22.5
Fannettsburg member:		
Calcarenite, light gray, medium to coarse	1.5	56.3
Calcarenite, medium gray, coarse, slabby; with a few thin beds of medium- to dark-gray, medium-grained, laminated limestone	8.7	54.8
Limestone, dark to medium gray, medium-grained to finely crystalline, laminated, platy to slabby; with a few thin interbeds of medium-gray calcarenite; contains *Solenopora*	17.6	46.1
Limestone, light to medium gray, medium-grained, platy, tan weathering	27.5	28.5
Calcarenite, medium gray, coarse	1.0	1.0
Pinesburg member:		
Limestone, dark gray, fine- to medium-grained, crinkly to slabby	70.5	89.3
Covered interval	14.8	18.8
Limestone, dark gray, fine- to medium-grained, cobbly	2.0	4.0
Calcarenite, medium to dark gray, coarse	2.0	2.0
"Stones River" limestone:		
Calcilutite, light to dark dove	—	—

Lower Middle Ordovician sections in the St. Thomas belts of outcrop, Franklin County, Pennsylvania

15. St. Thomas Section: Exposure on the south side of U. S. Route 30, 1.0 miles S. 70°W. of St. Thomas. Beds dip 50°, N. 80°E.

	Thickness (Feet) I	II
Mercersburg formation:		
Limestone, dark gray, fine-grained, crinkly bedded, blue weathering; interval partly covered	—	—
Limestone, dark gray, fine-grained, crinkly bedded, blue weathering	11.6	22.3
Limestone, dark to medium gray, coarse-grained to finely crystalline, slabby, gray weathering	10.7	10.7
Shippensburg formation:		
Doylesburg member:		
Limestone, dark gray to mottled dove, fine-grained, platy, blue weathering	1.0	12.0
Calcilutite, dove, massively bedded, white weathering; contains *Tetradium cellulosum* (Hall) and *Cryptophragmus*	11.0	11.0

	Thickness (Feet) I	II
Fannettsburg member:		
Limestone, dark gray, fine- to medium-grained, platy to slabby; contains *Cryptophragmus*	1.5	39.2
Limestone, medium gray, fine-grained to medium crystalline, slabby; with interbeds of light-gray, coarse, massively bedded calcarenite containing *Solenopora*	19.8	37.7
Calcarenite, medium gray, fine, laminated, slabby; with a few thin beds of medium-gray coarse calcarenite	15.5	17.9
Calcarenite, dark to medium gray, coarse slabby	2.4	2.4
Pinesburg member:		
Limestone, dark to medium gray to mottled dove, fine-grained, semicrystalline, slabby; contains *Opikina*, *Strophomena*	8.0	114.3
Covered interval	8.0	106.3
Limestone, dark gray, fine-grained, slabby, blue weathering; interval partly covered	20.4	98.3
Limestone, dark gray, cobbly; with prominent silty seams	5.8	77.9
Limestone, dark gray, fine-grained, cobbly to platy; with thick partings of buff-weathering siltstone	7.9	72.1
Metabentonite, shale and limestone unit; Shippensburg 1 (0.3 brown shale; 0.1 yellow metabentonite; 0.2 dark-gray limestone; 0.3 yellow brown shale; 0.1 lemon yellow metabentonite)	1.0	64.2
Limestone, dark gray, fine-grained, cobbly to platy; with thick partings of buff-weathering siltstone	10.3	63.2
Clay, yellow, gritty, shaly, probably metabentonitic; Shippensburg 1	0.1	52.9
Limestone, dark gray, fine-grained, cobbly, blue weathering; with prominent buff-weathering silty seams; contains *Sowerbyella* sp. cf. *S. aequistriata* (Willard) and *Nidulites pyriformis* Bassler	20.7	52.8
Limestone, dark gray to black, fine-grained, slabby, gray weathering	3.2	32.1
Limestone, dark gray, fine-grained, cobbly to platy; with thick partings of buff weathering siltstone; contains *Dinorthis transversa* Willard	9.3	28.9
Limestone, dark gray to black, medium-grained, platy, gray weathering; with irregular silty seams; slightly coarser-grained and darker weathering in bottom 2 feet	19.6	19.6
"Stones River" limestone:		
Calcilutite, light to dark dove, massively bedded, white weathering; contains *Tetradium cellulosum* ? (Hall)	—	—

16. Fort McCord Section: Compiled from exposures along stream and road 4.8 miles N. 20°E. of St. Thomas and 0.7 miles N. 50°W. of Freys. Beds dip 65°, S. 65°E.

	Thickness (Feet) I	II
Martinsburg shale:		
Limestone, black, fine-grained, argillaceous, shaly to slabby, buff weathering; contains *Cryptolithus* and graptolites	—	—
Metabentonite and shale unit; Salona 3 (?) (0.2 brown shale; 0.3 yellow metabentonite)	0.5	22.5
Limestone, black, fine-grained, argillaceous, shaly to slabby, buff weathering; interval partly covered	22.0	22.0
Oranda formation:		
Calcilutite, black, slabby, blue to white weathering; contains *Leptaena*, *Oxoplecia*, *Receptaculites*, and *Bimuria*	45.0	45.0
Lower part of formation covered	—	—
Mercersburg formation:		
Upper part of formation covered. Total thickness between exposures	13.0	—
Limestone, dark gray, fine- to-medium-grained; interval covered, but abundant float	26.8	63.1
Limestone, dark gray, fine- to medium-grained, crinkly bedded, blue weathering	30.5	36.3

	Thickness (Feet) I	II
Limestone, dark gray to black, medium-grained, semicrystalline, platy	4.8	5.8
Calcarenite, medium gray, coarse, platy; bottom contact is disconformable with 3 inches of relief	1.0	1.0
Ship ensburg formation:		
Doylesburg member:		
Limestone, dark dove, fine-grained, gray weathering; contains *Cryptophragmus*	1.5	20.3
Calcilutite, dark to light dove, massive, white weathering; contains *Cryptophragmus* and *Tetradium cellulosum* (Hall)	7.6	18.8
Conglomerate, fine pellets of dark-dove calcilutite; interbeds of dark-dove calcilutite	7.0	11.2
Limestone, medium gray, fine- to medium-grained, platy	2.1	4.2
Conglomerate, fine pellets of medium-gray, fine-grained limestone	2.1	2.1
Fannettsburg member:		
Limestone, medium gray, medium-grained, granular weathering; with a few interbeds of cystoidal calcarenite	7.1	56.7
Limestone, dark to medium gray, medium-grained, semicrystalline to finely crystalline, platy, cross-laminated	21.4	49.6
Covered interval	21.2	28.2
Calcarenite, medium gray, coarse, slabby	7.0	7.0
Pinesburg member:*		
Limestone, dark gray, fine-grained, cobbly to platy; with prominent silty seams	8.9	8.9
Lower part of member covered	—	—

17. Section at small quarry, 1.25 miles S. 18°W. of St. Thomas.

	Thickness (Feet) I	II
Mercersburg formation:		
Upper part of formation covered	—	—
Limestone, dark gray, fine-grained, crinkly bedded, blue weathering	4.9	9.2
Calcarenite, dark to medium gray, coarse, crinoidal	2.0	4.3
Limestone, dark gray, medium-grained, platy	1.9	2.3
Calcarenite, dark to medium gray, coarse; feathering out laterally; bottom contact is disconformable with 3 inches of relief	0.4	0.4
Shippensburg formation:		
Doylesburg member:		
Calcilutite, medium dove to tan, mottled, massive	2.6	16.1
Calcilutite, light dove, massive, white weathering	13.5	13.5
Fannettsburg member:		
Limestone, medium gray, medium-grained to coarsely crystalline, crinoidal	4.7	4.7
Lower part of member covered	—	—

18. Brandt Church Section: Compiled from field and road cut exposures, 2.7 miles S. 13°W. of St. Thomas. Beds dip 50°, S. 80°E.

	Thickness (Feet) I	II
Martinsburg shale:		
Shale, brown, papery to platy	15.2	198.1
Calcite, thin, white, platy, with a little yellow metabentonitic clay	0.1	182.9
Shale, brown, papery to platy	15.7	182.8
Limestone, black, fine-grained, argillaceous, buff weathering; in part weathers to pencil shale	19.5	167.1
Metabentonite, light yellow; with white calcite plates; Salona 7 (?)	0.1	147.6
Limestone, black, fine-grained, argillaceous, buff shaly weathering	9.8	147.5
Metabentonite and shale unit; Salona 6 (?) (0.1 brown shale; 0.1 deep-yellow metabentonite)	0.2	137.7

* Poor exposure and drift indicate that the Pinesburg member is approximately 70 feet thick, 0.5 miles south of the Fort McCord section.

	Thickness (Feet) I	II
Limestone, black, fine-grained, argillaceous, buff shaly weathering	47.2	137.5
Metabentonite and shale unit; (0.1 brown shale; 0.3 yellow shaly metabentonite with white calcite plates	0.4	90.3
Limestone, black, fine-grained, argillaceous, buff shaly weathering	16.9	89.9
Metabentonite, deep yellow; Salona 5 (?)	0.2	73.0
Limestone, black, fine-grained, argillaceous, buff shaly weathering	5.0	72.8
Metabentonite and shale unit; Salona 4 (?) (0.1 brownish-gray shale; 0.4 green to buff shaly metabentonite; 0.3 light-yellow and white metabentonite with white calcite plates)	0.8	67.8
Limestone, black, fine-grained, argillaceous, thick-bedded, buff weathering; with shaly calcareous interbeds	29.4	67.0
Metabentonite and shale unit; Salona 3 (?) (0.2 gray to brown papery; 0.3 buff to gray metabentonite)	0.5	37.6
Limestone, black, fine-grained, argillaceous, thick-bedded, buff weathering; with shaly calcareous interbeds	30.9	37.1
Metabentonite unit; Salona 2 (0.4 light-yellow and gray metabentonite; 0.1 deep-yellow shaly metabentonite; 0.4 green to buff shaly metabentonite)	0.9	6.2
Limestone, black, fine-grained, semicrystalline (crinoidal); weathers dark gray with purplish cast	5.3	5.3
Oranda formation:		
Calcilutite, black, slabby; with buff silty partings; contains *Oxoplecia* and *Receptaculites*	15.2	21.7
Metabentonite and shale unit; Salona 0 (0.1 gray paper shale; 0.2 light-yellow and white metabentonite)	0.3	6.5
Calcilutite, black, slabby; contains *Reuschella edsoni* (Bassler) and *Receptaculites*	4.7	6.2
Limestone, dark gray, medium-grained, semicrystalline, slabby	1.5	1.5
Mercersburg formation:		
Limestone, dark gray, fine- to medium-grained, thin and crinkly bedded, blue weathering	45.7	50.3
Calcarenite, dark gray, fine to coarse, slabby	4.6	4.6
Shippensburg formation:		
Doylesburg member:		
Limestone, dark to dove gray, fine- to medium-grained, platy; with a few interbeds of dark-dove to tan, calcilutite-pellet conglomerate; contains abundant *Cryptophragmus*	16.6	20.8
Conglomerate, dark gray to dove, calcilutite-pellet; pellets up to 1 inch in diameter	4.2	4.2
Fannettsburg member:		
Limestone, dark to medium gray, medium-grained, laminated, platy; with a few beds of coarse calcarenite containing *Glyptorthis* and *Solenopora*; lower 20 feet partly covered	30.4	30.8
Calcarenite, medium gray, medium-textured	0.4	0.4
Pinesburg member:		
Limestone, dark gray, fine- to medium-grained, platy, blue weathering	5.8	111.9
Covered interval	51.2	106.1
Limestone, dark gray, fine-grained, cobbly to platy, blue weathering	54.9	54.9
Base of member covered	—	—
"Stones River" limestone:		
Upper part of formation covered. Total thickness between exposures	11.7	—
Calcilutite, medium gray to dove, white weathering; contains *Tetradium syringoporoides* Ulrich; interval partly covered	11.9	11.9

19. Rockdale Section: Exposure along Cumberland Valley Railroad, 0.5 miles N. 45°W. of Williamson.

	Thickness (Feet) I	II
Oranda formation:		
Upper part of formation covered	—	—
Calcilutite, black, slabby; interval mostly covered	3.0±	3.8±
Calcarenite, dark gray, coarse	0.8±	0.8±
Mercersburg formation:		
Limestone, dark to medium gray, fine-grained, slabby; upper 14 feet partly covered	25.2	69.0
Metabentonite, light yellow; with white calcite plates	0.3	43.8
Limestone, dark gray, fine-grained, thin and crinkly bedded, blue weathering	30.2	43.5
Calcarenite, dark gray, fine- to medium-textured, slabby	13.3	13.3
Shippensburg formation:		
Doylesburg member:		
Limestone, light to medium gray, fine- and medium-grained, slabby; with a 3-inch bed of dark-gray, fine-grained, limestone-pellet conglomerate at top	5.0	6.0
Conglomerate of medium-gray, fine- to medium-grained, limestone pellets up to 3 inches in diameter	1.0	1.0
Fannettsburg member:		
Limestone, dark gray, medium-grained to medium crystalline, faintly laminated, slabby	4.5	19.2
Calcarenite, medium gray, coarse-textured (crinoidal); contains *Glyptorthis*	0.5	14.7
Limestone, dark gray, fine- to medium-grained, thick-bedded; with a few faintly laminated granular beds and a 3-inch crinoidal calcarenite at bottom; lower 7 feet partly covered	14.2	14.2
Pinesburg member:		
Limestone, dark gray, fine-grained, thick-bedded, blue weathering; interval partly covered	25.0	38.0
Covered interval	11.5	13.0
Limestone, medium gray, fine-grained, thin-bedded; with prominent buff-weathering silty partings	1.5	1.5
Lower part of member covered	—	—

Lower Middle Ordovician sections in the Pinesburg-Wilson belt of outcrop, Washington County, Maryland

20. Wilson section: Exposure along U. S. Route 40 at Conococheague Creek, 4.3 miles N. 20°W. of Williamsport, Md.

	Thickness (Feet) I	II
Mercersburg formation:		
Upper part of formation covered	—	—
Metabentonite, light yellow to white, thick; with a little float of buff to cream shale	—	—
Limestone, dark gray, fine- to medium-grained, crinkly	11.2	46.9
Limestone, dark gray, fine- to medium-grained, platy to crinkly	14.4	35.7
Limestone, dark gray, fine-grained, slabby; with prominent shale partings	5.5	21.3
Metabentonite unit; Mercersburg 2 (upper) (0.1 yellow shaly metabentonite; 0.4 white to light-yellow metabentonite with white calcite plates)	0.5	15.8
Limestone, dark gray, fine-grained, slabby; contains *Nidulites*; top bedding surface is black and silicified	10.9	15.3
Limestone, dark gray, fine- to medium-grained, crinkly to platy	2.4	4.4
Metabentonite and shale unit; Mercersburg 2 (lower) (1.0 gray, medium-grained, calcareous, laminated, spheroidally weathering siltstone with brown shale interbeds; 0.5 brown shale; 0.2 yellow and white metabentonite with white calcite plates; 0.2 brown shale; 0.1 yellow-brown shaly metabentonite)	2.0	2.0

	Thickness (Feet) I	II
Shippensburg formation:		
Fannettsburg member:		
Limestone, dark gray, fine-grained, slabby, gray to buff weathering; top bedding surface is black and silicified	11.5	26.9
Limestone, dark gray, platy, buff papery weathering; with prominent shaly partings	1.0	15.4
Metabentonite and shale unit; Shippensburg 6 (0.1 olive drab shale, 0.1 light-yellow metabentonite with white calcite plates)	0.2	14.4
Limestone, dark gray, fine-grained, slabby; contains *Nidulites*	7.6	14.2
Metabentonite and shale unit; Shippensburg 5 (0.1 brown paper shale; 0.2 light-yellow to white shaly metabentonite with white calcite plates)	0.3	6.6
Limestone, dark gray, fine-grained, slabby; contains *Nidulites*	4.1	6.3
Limestone, platy, blue to buff weathering; with thick brown shale interbeds	0.5	2.2
Metabentonite and shale unit; Shippensburg 4 (0.1 brown fissile shale; 0.1 light-yellow to white metabentonite with white calcite plates)	0.2	1.7
Limestone, dark gray, crinkly	1.1	1.5
Calcarenite, medium gray, coarse, laminated	0.4	0.4
Pinesburg member:		
Limestone, dark gray, fine- to medium-grained, thin crinkly bedded	16.3	238.7
Metabentonite and shale unit; Shippensburg 3 (0.1 brown shale; 0.1 yellow and white metabentonite with white calcite plates)	0.2	222.4
Limestone, dark gray, fine- to medium-grained, thin and crinkly bedded; contains *Lambeophyllum*	23.0	222.2
Limestone, dark gray, fine-grained, slabby	53.8	199.2
Limestone, dark gray, fine- to medium-grained, thin and crinkly bedded	6.5	145.4
Limestone, dark gray, fine-grained, slabby to massive; contains *Lambeophyllum*, *Diplotrypa*, and *Nidulites*	47.0	138.9
Metabentonite, medium yellow to white; Shippensburg 2	0.2	91.9
Limestone, dark gray, fine-grained, slabby	21.7	91.7
Limestone, dark gray, fine-grained, platy; contains *Nidulites*, *Isotelus*	9.7	70.0
Limestone, dark gray, fine-grained, cobbly to platy	8.6	60.3
Clay, yellow, silty, metabentonitic; Shippensburg 1	0.1	51.7
Limestone, dark gray, fine-grained, cobbly to platy; contains *Dinorthis transversa Willard*	5.9	51.6
Clay and shale unit; Shippensburg 1 (0.2 yellow-brown shale; 0.1 yellow, silty, metabentonitic clay)	0.3	45.7
Limestone, dark gray, fine-grained, platy; contains *Dinorthis transversa* Willard, *Isotelus*	2.1	45.4
Clay and shale unit; Shippensburg 1 (0.1 brown shale; 0.1 yellow, shaly, metabentonitic clay)	0.2	43.3
Limestone, dark gray, fine-grained, cobbly; with shaly partings	3.1	43.1
Shale, yellow brown, metabentonitic; Shippensburg 1	0.1	40.0
Limestone, dark gray, fine-grained, cobbly; with shaly partings	3.8	39.9
Clay and shale unit; Shippensburg 1 (0.2 olive-drab shale; 0.1 yellow, silty, metabentonitic clay)	0.3	36.1
Limestone, dark gray, fine-grained, cobbly; with shale partings	4.0	35.8
Metabentonite, yellow, shaly; Shippensburg 1	0.1	31.8
Limestone, dark gray, fine-grained, cobbly; with shaly partings	1.8	31.7
Limestone, dark gray, fine-grained, semicrystalline, slabby; contains abundant *Echinosphaerites*	1.0	29.9
Limestone, dark gray, fine-grained, cobbly; with prominent silty seams; contains *Echinosphaerites*	5.4	28.9
Limestone, dark gray, fine- to medium-grained, platy to cobbly; contains *Echinosphaerites* and *Dinorthis transversa* Willard	10.2	23.5
Covered interval	3.4	13.3
Limestone, black, fine- to medium-grained, cobbly, brown weathering; with argillaceous partings	9.9	9.9
"Stones River" limestone:		
Calcilutite, light to dark dove, massive; contains *Tetradium cellulosum* (Hall)	—	—

21. Pinesburg Station Section: Exposure along Western Maryland Railroad, 2.7 miles N. 64°W. of Williamsport, Md. Beds dip 75°, S. 80°E.

	Thickness (Feet) I	II
Shippensburg formation:		
Pinesburg member:		
Scattered field exposure of thick-bedded, *Nidulites*-bearing limestone	—	—
Limestone, dark gray, fine-grained, platy	9.0	101.8
Limestone, dark gray, fine-grained, slabby; contains *Nidulites*	4.4	92.8
Limestone, dark gray, fine-grained, platy; contains *Nidulites*	7.2	88.4
Limestone, dark gray, fine-grained, cobbly; with prominent silty partings	18.8	81.2
Metabentonite and shale, unit; Shippensburg 1 (0.1 yellow metabentonite; 0.1 brownish-yellow metabenonitic shale; 0.1 yellow and white metabentonite)	0.3	62.4
Limestone, dark gray, fine-grained, cobbly; with prominent silty partings	1.2	62.1
Clay, yellow, metabentonitic, shaly; Shippensburg 1	0.1	60.9
Limestone, dark gray; fine-grained, cobbly; with prominent silty partings	2.1	60.8
Shale, yellow, metabentonitic; Shippensburg 1	0.1	58.7
Limestone, dark gray, fine-grained, cobbly; with prominent silty seams	2.2	58.6
Clay, yellow, silty, metabentonitic; Shippensburg 1	0.3	56.4
Limestone, dark gray, fine-grained, cobbly; with prominent silty seams; contains abundant *Echinosphaerites* at bottom	7.4	56.1
Shale, yellow, metabentonitic; Shippensburg 1	0.2	48.7
Limestone, dark gray, fine-grained, cobbly; with prominent silty seams	2.2	48.5
Metabentonite and shale unit; Shippensburg 1 (0.3 yellow metabentonitic shale; 0.1 light-yellow metabentonite with white calcite plates)	0.4	46.3
Limestone, dark gray, fine-grained, thick-bedded; contains *Echinosphaerites* at top	12.1	45.9
Limestone, dark gray, fine-grained, platy to crinkly	9.9	33.8
Clay, light yellow to orange, possibly metabentonitic	0.3	23.9
Limestone, dark gray, fine-grained, platy; contains abundant *Echinosphaerites* in upper half	14.3	23.6
Limestone, dark gray, fine-grained, cobbly; with prominent gray to buff-weathering silty partings	9.3	9.3
"Stones River" limestone:		
Calcilutite, dark to light dove, thick-bedded	—	—

Lower Middle Ordovician sections in the Middle Spring-Newville belt of outcrop, Franklin and Cumberland Counties, Pennsylvania

22. Elliotts Mill Section: Compiled from exposure in small quarry and along secondary road at Conodoguinet Creek, 0.9 miles N. 42°E. of Plainfield. Beds dip 65°, N. 30°W.

	Thickness (Feet) I	II
Oranda formation:		
Upper part of formation covered	—	—
Limestone, dark gray to black, platy to slabby; contains *Oxoplecia, Christiania, Illaenus*	28.0	137.9
Limestone, black, fine-grained, slabby; lower 67 feet partly covered	106.9	109.9
Limestone, dark gray, medium-grained, semicrystalline, slabby, dark-gray weathering; upper part of interval covered	3.0	3.0
Mercersburg formation:		
Limestone, dark gray, fine-grained, crinkly to platy, light-gray weathering; contains *Diplotrypa*	66.6	108.1
Limestone, dark gray, fine-grained, platy to slabby	33.3	41.5
Metabentonite and siltstone unit; Mercersburg 2 (0.2 olive-drab shale and siltstone; 0.6 yellow metabentonite with white calcite plates)	0.8	8.2

	Thickness (Feet) I	II
Limestone, dark gray, fine-grained, cobbly to platy; contains *Illaenus* and *Sowerbyella*	7.4	7.4
Lower part of formation covered	—	—
Shippensburg formation:		
Pinesburg member:		
Upper part of member covered. Total thickness between exposures	62.0	—
Limestone, dark gray, fine- to medium-grained, slabby to crinkly	47.0	203.2
Limestone, dark gray, fine- to medium-grained, cobbly to platy; with prominent silty partings	90.7	156.2
Limestone, dark gray, fine- to medium-grained, slabby	20.1	65.5
Limestone, dark gray, fine-grained, cobbly, blue weathering	32.0	45.4
Limestone, dark gray to black, fine-grained, platy to cobbly	11.4	13.4
Limestone, dark gray, medium-grained to finely crystalline, platy	2.0	2.0
"Stones River" limestone:		
Limestone, dark gray to dove, fine-grained, crinkly to massive, blue to white weathering	—	—

23. **Diller Church Section:** Compiled from exposure along the Pennsylvania Turnpike, 2.9 miles N. 45°E. of Newville. Beds dip 80°, S. 50°E.

	Thickness (Feet) I	II
Martinsburg shale:		
Covered interval	—	—
Limestone, fine-grained, argillaceous, laminated, platy, white to buff weathering	9.0	16.0
Covered interval	3.4	7.0
Metabentonite, lemon yellow; Salona 2	0.1	3.6
Limestone, dark gray, fine-grained, argillaceous, platy to shaly, brown weathering; with a coarse calcarenite bed at top; contains *Cryptolithus*, *Sinuites* and *Lingula*	3.5	3.5
Oranda formation:		
Limestone, black, fine-grained, semicrystalline, slabby, whitish weathering; contains *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens). *Receptaculites*, *Cryptolithus*	3.4	148.7
Metabentonite, white; Salona 0	0.3	145.3
Limestone, black, fine-grained, slabby, purplish weathering	4.2	145.0
Covered interval	27.5	140.8
Limestone, dark gray to black, fine-grained, massive; contains *Oxoplecia simulatrix*; interval partly covered	56.0	113.3
Covered interval	14.0	57.3
Limestone, black, fine-grained, massive to slabby	26.0	43.3
Covered interval	11.0	17.3
Metabentonite, cream to yellow	0.4	6.3
Covered interval	2.8	5.9
Limestone, black, fine- to medium-grained, semicrystalline, crinkly to slabby, purplish weathering	3.1	3.1
Mercersburg formation:		
Kauffman member:		
Limestone, dark gray, fine-grained, platy to crinkly, blue weathering; contains *Diplotrypa*	17.4	76.7
Covered interval	7.7	59.3
Limestone, dark gray, fine- to medium-grained, platy to slabby, blue weathering; contains *Nidulites*, *Leptaena* sp. cf. *L. charlottae* Winchell and Schuchert, *Diplotrypa*	19.5	51.6
Metabentonite, yellow, with white calcite plates; Mercersburg 3	0.3	32.1
Limestone, dark gray, fine-grained, crinkly to platy; contains *Nidulites*	30.8	31.8
Metabentonite and siltstone unit; Mercersburg 2 (0.4 olive-drab shale and siltstone; 0.6 light-yellow metabentonite)	1.0	1.0
Housum member:		
Limestone, dark to medium gray, fine-grained, crinkly, gray weathering	1.0	53.5

	Thickness (Feet) I	II
Covered interval	8.7	52.5
Limestone, dark gray, fine- to medium-grained, crinkly to cobbly; with prominent silty partings	21.0	43.8
Limestone, dark dove gray, fine-grained, slabby, white to buff weathering; contains *Leptaena* sp. cf. *L. Charlottae* Winchell and Schuchert; lower 10 feet poorly exposed	19.6	22.8
Metabentonite, light-yellow and white; Mercersburg 1	0.1	3.2
Limestone, dark gray, fine-grained, thin platy; with a few interbeds of dark-gray, medium-grained, semicrystalline limestone at bottom	3.1	3.1
Shippensburg formation:		
Fannettsburg member:		
Limestone, dark gray, fine-grained, slabby, whitish weathering	5.4	6.7
Limestone, dark gray, fine-grained, slabby, buff papery weathering; with prominent shale partings	1.0	1.3
Metabentonite, yellow; Shippensburg 4	0.3	0.3
Lower part of member covered	—	—
Pinesburg member:		
Upper part of member covered. Total thickness between exposures	10.8	—
Limestone, dark gray, fine-grained, crinkly to slabby; contains *Lambeophyllum* and *Nidulites*	24.0	174.0
Limestone, dark gray, fine-grained, platy; with silty partings; contains *Nidulites*	18.0	150.0
Covered interval	17.0	132.0
Limestone, dark gray, fine-grained, cobbly; with prominent silty partings	9.0	115.0
Covered interval	10.0	106.0
Limestone, dark gray, fine-grained, cobbly; with prominent silty partings	6.3	96.0
Limestone, dark gray, medium-grained, semicrystalline	0.7	89.7
Covered interval	55.0	89.0
Limestone, dark gray, fine-grained, argillaceous, gray weathering	34.0	34.0
Lower part of formation covered	—	—

24. Kough Quarry Section: Western environs of Newville, Pennsylvania.

	Thickness (Feet) I	II
Mercersburg formation:		
Housum member:		
Upper part of member covered	—	—
Limestone, dark gray, fine-grained, slabby; with prominent shaly partings	4.2	8.3
Metabentonite, white and light yellow; with abundant white calcite plates; Mercersburg 1	0.3	4.1
Limestone, dark gray, fine-grained, slabby; with prominent shaly partings; contains *Nidulites*; a layer of black chert nodules above a thick shale parting at bottom	3.8	3.8
Shippensburg formation:		
Fannettsburg member:		
Limestone, dark gray, fine-grained, slabby	3.7	5.0
Limestone, tan to gray, fine-grained, slabby, buff papery weathering; with thick shaly partings	0.9	1.3
Metabentonite unit; Shippensburg 4 (0.2 yellow shaly metabentonite; 0.2 light-yellow to white metabentonite)	0.4	0.4
Pinesburg member:		
Limestone, dark gray, fine-grained, slabby; with thin shaly partings	4.0	167.3
Limestone, dark gray, fine- to medium-grained, crinkly, blue-gray weathering	8.7	163.3
Metabentonite, cream to lightyellow; with white calcite plates; Shippensburg 3	0.2	154.6
Limestone, dark gray, fine- to medium-grained, crinkly, blue-gray weathering	33.0	154.4

	Thickness (Feet) I	II
Limestone, dark to medium gray, fine-grained, cobbly to platy; with prominent partings	6.0	121.4
Metabentonite, yellow; with white calcite plates; Shippensburg 2	0.3	115.4
Limestone, dark to medium gray, fine-grained, cobbly to platy; with prominent partings	25.0	115.1
Limestone, dark gray, fine-grained, thin cobbly; with irregular silty partings; contains *Nidulites*	23.0	90.1
Limestone, dark gray, medium-grained, gray weathering; with prominent partings	4.3	67.1
Limestone, dark gray, fine-grained, platy, blue weathering	1.0	62.8
Limestone, dark gray, fine- to medium-grained, blue weathering; with a few interbeds of medium calcarenite	4.3	61.8
Calcarenite, dark gray, medium to coarse, slabby gray granular weathering	9.7	57.5
Limestone, dark gray, fine- to medium-grained, cobbly; with prominent silty partings; contains *Dinorthis transversa* Willard and *Sowerbyella* sp. cf. *S. aequistriata* Willard	31.0	47.8
Limestone, dark gray, medium-grained, semicrystalline, platy; with dark-gray silty partings	16.8	16.8
"Stones River" limestone:		
Calcilutite, dove, massive; with buff-weathering interbeds; contains *Tetradium syringoporoides* Ulrich; 2.2 feet of medium to coarse calcarenite at top	—	—

25. Oakville Section: Compiled from exposure along secondary road, 1 mile N. 52°W. of Oakville.

	Thickness (Feet) I	II
Martinsburg shale:		
Metabentonite, yellow	0.3	87.2
Limestone, black, fine-grained, argillaceous, platy to shaly, buff weathering	56.9	86.9
Metabentonite, light gray; with abundant white calcite plates	0.1	30.0
Limestone, black, fine-grained, argillaceous, platy to shaly, buff weathering	0.9	29.9
Metabentonite, yellow; with abundant plates of white calcite	0.1	29.0
Limestone, black, fine-grained, argillaceous, shaly to platy, buff weathering	9.7	28.9
Shale, brown, pencil	1.5	19.2
Limestone, black, argillaceous, platy, buff weathering	2.1	17.7
Metabentonite unit; Salona 3 (?) (0.1 cream-colored metabentonite; 0.2 yellow shaly metabentonite; 0.1 cream-colored metabentonite)	0.4	15.6
Limestone, black, argillaceous, platy, buff weathering	7.8	15.2
Limestone, black to brown, fine-grained, argillaceous, platy to slabby, purplish weathering	2.1	7.4
Metabentonite and shale unit; Salona 2 (0.1 brown shale; 0.1 light-yellow shaly metabentonite; 0.1 light-yellow metabentonite)	0.3	5.3
Limestone, black to brown, fine-grained, argillaceous, platy to slabby, purplish weathering; contains *Cryptolithus* and *Sinuites*	5.0	5.0
Oranda formation:		
Limestone, black, fine-grained, massive, blue to whitish weathering, contains *Cryptolithus*	33.3	79.2
Metabentonite unit; (0.2 yellow shaly metabentonite; 0.1 yellow metabentonite with plates of white calcite)	0.3	45.9
Limestone, black, fine-grained, massive, blue to whitish weathering; contains *Leptaena* sp. aff. *L. rhomboidalis* (Wilckens)	27.0	45.6
Fault, displacement small but unknown	—	—
Limestone, dark gray to black, fine-grained, massive; fractured and brecciated in lower part	18.6	18.6
Covered interval	—	—

Poorly exposed interval representing about 550 feet of beds. Exposure of Shippensburg metabentonites 3, 4, and 5 indicates structural repetition of at least part of the section.

	Thickness (Feet) I	II
Mercersburg formation:		
Kauffman member:		
Covered interval	—	—
Limestone, dark gray, fine- to medium-grained, dark-gray weathering; upper 17 feet poorly exposed	39.3	50.6
Metabentonite and siltstone unit; Mercersburg 2 (0.3 yellow and white metabentonite; 1.0 dark-gray, massive limestone; 8.0 olive-drab siltstone float; 2.0 yellow and white metabentonite)	11.3	11.3
Housum member:		
Covered interval	59.1	62.1
Shale, yellow brown, silty, poorly exposed. Probably represents Mercersburg metabentonite 1	3.0	3.0
Shippensburg formation:		
Fannettsburg member:		
Limestone, dark gray, fine-grained, platy; with prominent partings	23.5	24.5
Clay, yellow brown, metabentonitic; Shippensburg 5 (?)	1.0	1.0
Covered interval	—	—
Pinesburg member:		
Covered interval. Total thickness between exposures	49.3	—
Limestone, dark gray, fine-grained, cobbly; with prominent silty seams	8.9	136.1
Metabentonite, yellow; Shippensburg 2	0.2	127.2
Limestone, dark gray, fine-grained, cobbly; with prominent silty seams	27.4	127.0
Covered interval	58.4	99.6
Limestone, dark gray, medium-grained, argillaceous; with prominent silty seams	41.2	41.2
Lower part of member covered	—	—
"Stones River" limestone:		
Upper part of member covered. Total thickness between exposures	14.0	—
Calcilutite, dove, massive; scattered field exposure	—	—

26. Middle Spring Section: Compiled from exposures along Pennsylvania Route 696 and in fields to south, 2.5 miles N. 31°W. of Shippensburg, Pennsylvania. Beds dip 70°, S. 50°E.

	Thickness (Feet) I	II
Oranda formation:		
Limestone, black, fine-grained, slabby; interval largely covered	7.5	45.1
Covered interval	5.0	37.6
Limestone, dark gray to black, fine-grained, slabby; contains *Cryptolithus*	21.6	32.6
Limestone, dark gray to black, medium-grained, platy	3.2	11.0
Limestone, dark gray, medium-grained, crinkly to platy, gray granular weathering; with lens of calcarenite at top	6.2	7.8
Metabentonite, yellow	0.2	1.6
Limestone, dark gray, medium-grained, crinkly to platy, gray granular weathering	1.4	1.4
Mercersburg formation:		
Kauffman member:		
Limestone, dark gray, fine-grained, crinkly bedded, blue weathering	6.5	105.8
Metabentonite, light yellow; Mercersburg 4	0.3	99.3
Limestone, dark gray, fine-grained, crinkly bedded, blue weathering; contains *Nidulites* and *Diplotrypa*	14.6	99.0
Limestone, dark gray, fine- to medium-grained, platy; upper 15 feet of interval partly covered	26.1	84.4
Limestone, dark gray, medium-grained, crinkly bedded, highly fractured; interval partly covered	16.0	58.3
Limestone, dark gray, medium-grained, platy to crinkly, gray granular weathering; contains *Girvanella*	27.0	42.3
Limestone, dark gray to black, medium-grained	5.8	15.3
Metabentonite and siltstone unit; Mercersburg 2 (7.9 medium-gray, medium-grained, thick-bedded, olive-drab weathering calcareous		

	Thickness (Feet) I	II
siltstone with thick interbeds of greenish blocky siltstone and shale; 0.6 greenish-gray shale; 0.8 cream to greenish, metabentonitic shale; 0.2 light-yellow shaly metabentonite)	9.5	9.5
Housum member:		
Limestone, dark gray, fine-grained, platy to cobbly; with prominent silty partings; upper 2 inches black and silicified	17.0	49.2
Limestone, dark dove to tan, slabby, buff weathering; contains *Nidulites*	30.3	32.2
Metabentonite and siltstone unit; Mercersburg 1 (1.6 olive-drab, blocky to shaly siltstone; 0.3 light-yellow shaly metabentonite)	1.9	1.9
Shippensburg formation:		
Fannettsburg member:		
Limestone, dark gray, fine-grained, slabby; with prominent silty partings; top surface black and silicified; interval partly covered	30.5	36.8
Metabentonite, light yellow; with white calcite plates; Shippensburg 5	0.3	6.3
Limestone, dark gray, fine- to medium-grained, crinkly bedded; contains black chert nodules in upper 6 inches	2.5	6.0
Covered interval	1.5	3.5
Limestone, medium gray to tan, fine-grained, argillaceous, buff papery weathering	1.7	2.0
Metabentonite, yellow; with white calcite plates; Shippensburg 4	0.3	0.3
Pinesburg member:		
Limestone, dark gray, fine-grained, platy to slabby	47.0	188.0
Limestone, dark gray, fine- to medium-grained, platy to cobbly; with prominent buff-weathering silty seams; contains *Strophomena, Nidulites, Diplotrypa*	94.7	141.0
Metabentonite and shale unit; Shippensburg 1 (0.2 greenish shale; 0.1 yellow metabentonite)	0.3	46.3
Limestone, dark gray, medium-grained, platy to cobbly, gray weathering; with prominent argillaceous partings, contains *Dinorthis transversa* Willard	46.0	46.0
"Stones River" limestone:		
Calcilutite, dark to light dove, massive, white weathering, contains *Tetradium syringoporoides* Ulrich	—	—

Lower Middle Ordovician section in the Chambersburg belt of outcrop, Franklin County, Pennsylvania

27. Marion Section: "Type" section, compiled from exposure along the Cumberland Valley Railroad and in adjacent fields, 2.1 miles S. 53°W. of Marion. Beds dip 55°, N. 85°W.

	Thickness (Feet) I	II
Martinsburg shale:		
Covered interval	—	—
Limestone, dark gray, fine-grained, argillaceous, buff weathering; float only	—	—
Limestone, light gray to black, semicrystalline to coarsely crystalline, platy; contains *Sinuites, Lingula* and *Cryptolithus*	5.0	10.0
Metabentonite and siltstone unit; olive-drab blocky to shaly siltstone with light-yellow metabentonite at base; poorly exposed; Salona 2	4.0	5.0
Limestone, medium gray, medium-grained to medium crystalline; poorly exposed	1.0	1.0
Oranda formation:		
Limestone, dark gray to black, fine-grained, slabby; contains *Christiania*; interval partly covered	17.5	52.2
Metabentonite, light yellow; Salona 0	0.7	34.7
Limestone, dark gray to black, fine-grained, slabby; contains abundant *Echinosphaerites* at bottom; interval partly covered	30.0	34.0
Calcarenite, medium gray, medium to coarse, slabby; contains *Echinosphaerites*	4.0	4.0

	Thickness (Feet) I	II
Mercersburg formation:		
Kauffman member:		
Limestone, dark gray, fine- to medium-grained, crinkly to platy, blue-gray weathering	36.6	168.1
Metabentonite and shale unit; Mercersburg 4 (1.2 greenish shale with thin interbeds of gray, silty limestone; 0.3 yellow metabentonite)	1.5	131.5
Limestone, dark gray, fine- to medium-grained, crinkly to platy, blue-gray weathering; contains *Nidulites*; top bedding surface silicified	44.5	130.0
Metabentonite and shale unit; Mercersburg 3 (1.8 cream to gray-green, platy shale; 0.2 yellow metabentonite)	2.0	85.5
Limestone, dark gray, fine- to medium-grained, crinkly to platy, blue-gray weathering; top bedding surface silicified; interval partly covered	66.0	83.5
Metabentonite and shale unit; Mercersburg 2 (upper) (2.0 olive-drab blocky to platy shale; 0.5 greenish-yellow shaly metabentonite)	2.5	17.5
Limestone, dark gray, fine- to medium-grained, crinkly to platy	9.8	15.0
Limestone, medium gray, medium-grained to medium crystalline, slabby	1.4	5.2
Metabentonite and shale unit; Mercersburg 2 (lower) (2.6 olive-drab blocky, calcareous siltstone; 1.1 cream to green blocky shale; 0.1 cream to light-yellow metabentonite)	3.8	3.8
Housum member:		
Limestone, tan to dove gray, fine-grained, platy, buff weathering	16.0	67.0
Limestone, medium gray, fine- to medium-grained, crinkly to platy, gray weathering	10.5	51.0
Limestone, tan to dove gray, fine-grained, slabby, buff weathering	40.5	40.5
Shippensburg formation:		
Fannettsburg member:		
Limestone, dark gray, fine- to medium-grained, platy; with pocked silicified bedding surfaces at 5.0 feet and 9.5 feet in unit	9.5	262.8
Limestone, dark gray, fine-grained, massive, blue weathering; with thin, even, silty partings	96.3	253.3
Limestone, dark gray, fine-grained, irregular slabby, blue weathering; contains *Carabocrinus*, *Caryocystites*	8.6	157.0
Metabentonite and shale unit; Shippensburg 7 (0.2 yellow to gray, clay shale; 0.2 yellow metabentonite)	0.4	148.4
Limestone, dark gray, fine-grained, irregular slabby, blue weathering; with buff silty partings	1.4	148.0
Limestone, dark gray, fine- to medium-grained, crinkly, blue-gray weathering	18.0	146.6
Limestone, tan to light gray, fine-grained, papery to dense, white papery weathering; grades upward to dark-gray, fine-grained limestone	3.8	128.6
Metabentonite and shale unit; Shippensburg 6 (0.2 gray clay shale; 0.2 light-yellow metabentonite with white calcite plates)	0.4	124.8
Limestone, dark gray, fine-grained, slabby, blue weathering	5.4	124.4
Limestone, dark gray, fine- to medium-grained, crinkly, blue-gray weathering	5.9	119.0
Limestone, dark gray, fine-grained, slabby, blue weathering	7.1	113.1
Limestone, dark gray, fine- to medium-grained, crinkly, stylolitic	15.3	106.0
Limestone, dark gray, fine- to medium-grained, slabby	14.7	90.7
Limestone, dark gray, fine- to medium-grained, crinkly, stylolitic	7.9	76.0
Limestone, dark gray, fine-grained, massive, blue weathering	20.3	68.1
Limestone, dark gray, fine-grained, crinkly; contains *Nidulites*	4.7	47.8
Limestone, dark gray, fine-grained, slabby; lower part covered	26.2	43.1
Limestone, dark gray, fine- to medium-grained, crinkly, gray weathering, crinoidal	10.3	16.9
Limestone, medium gray, argillaceous, platy, buff weathering; with prominent shale partings	0.8	6.6
Metabentonite and shale unit; Shippensburg 4 (0.5 gray-green shale; 0.2 green to brown clay shale; 0.1 yellow metabentonite with white calcite plates)	0.8	5.8

	Thickness (Feet) I	II
Limestone, dark to medium gray, fine-grained, slabby, blue weathering; with two interbeds of light to medium-gray, medium-grained to medium-crystalline limestone at bottom; contains *Glyptorthis*	5.0	5.0
Pinesburg member:		
Limestone, dark gray, fine- to medium-grained, crinkly, stylolitic, blue weathering	10.0	152.9
Limestone, dark to medium gray, fine-grained, slabby, blue-gray weathering	20.0	142.9
Limestone, dark to medium gray, fine-grained, platy to slabby; with prominent silty partings; contains *Echinosphaerites*	23.1	122.9
Limestone, dark gray, fine- to medium-grained, cobbly to platy; contains *Dinorthis transversa* Willard	12.0	99.8
Clay and shale unit; Shippensburg 1 (0.2 gray clay shale; 0.1 yellow metabentonitic clay; 0.2 gray clay shale; 0.1 yellow metabentonitic clay)	0.6	87.8
Limestone, dark gray, fine- to medium-grained, cobbly; with prominent shaly seams	6.2	87.2
Clay, yellow, metabentonitic; Shippensburg 1	0.1	81.0
Limestone, dark gray, fine- to medium-grained, cobbly; with prominent shaly partings	0.7	80.9
Clay and shale unit; Shippensburg 1 (0.1 gray clay shale; 0.1 yellow metabentonitic clay)	0.2	80.2
Limestone, dark gray, fine- to medium-grained, cobbly; with prominent shaly partings; contains *Echinosphaerites*	10.2	80.0
Limestone, dark gray, fine- to medium-grained, platy to cobbly; with thin silty partings	8.7	69.8
Limestone, dark gray, fine-grained, cobbly; with prominent silty partings	14.3	61.1
Limestone, dark gray, fine-grained, platy, blue weathering	10.9	46.8
Limestone, dark gray, fine- to medium-grained, cobbly to slabby; with prominent silty partings; contains *Dinorthis transversa* Willard and *Sowerbyella* sp. cf. *S. aequistriata* Willard; lower part of member poorly exposed	35.9	35.9
"Stones River" limestone:		
Calcilutite, dove, massive, white weathering; and medium-gray, fine-grained, blue-weathering limestone; upper part of formation poorly exposed	—	—

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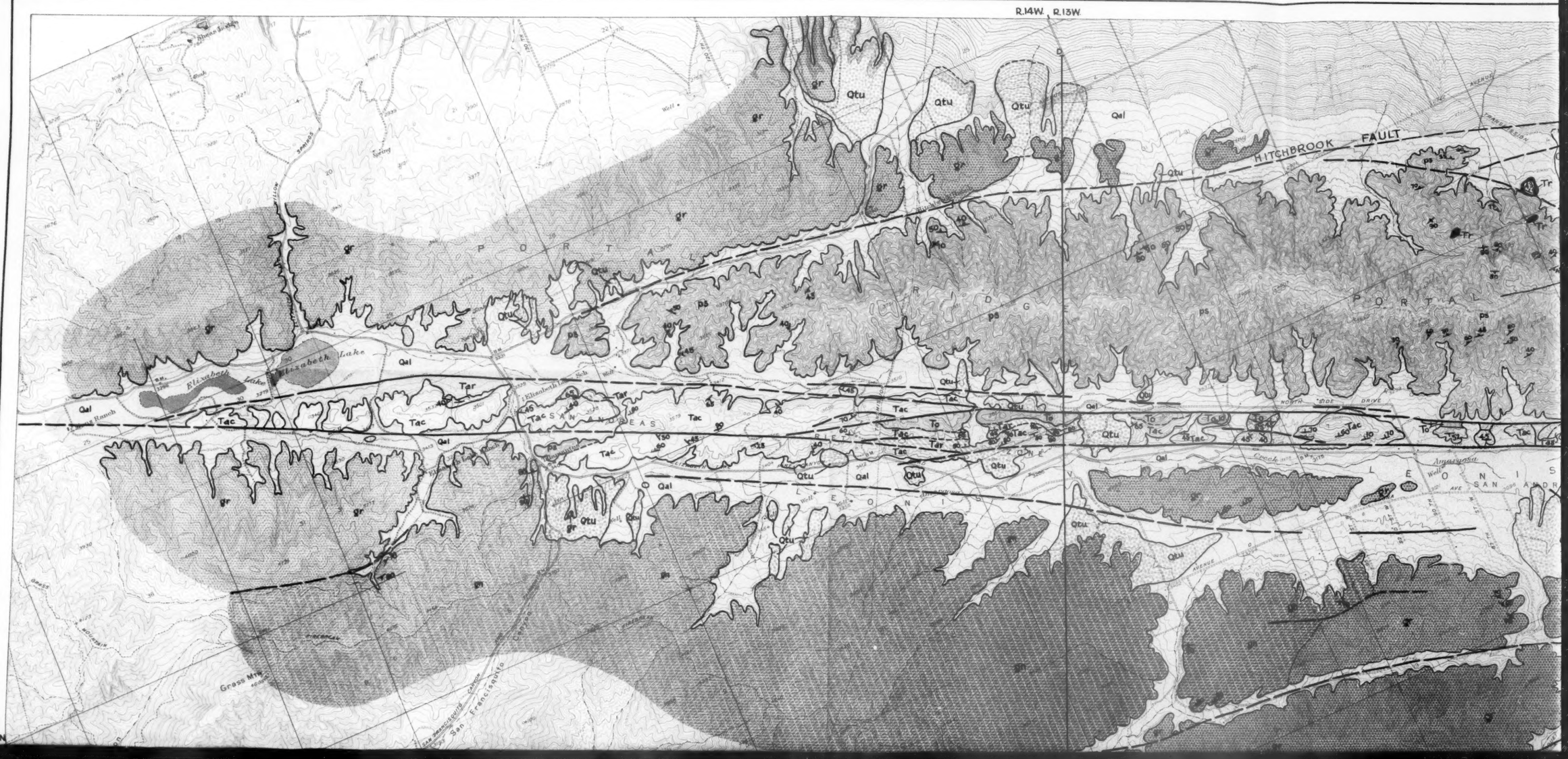
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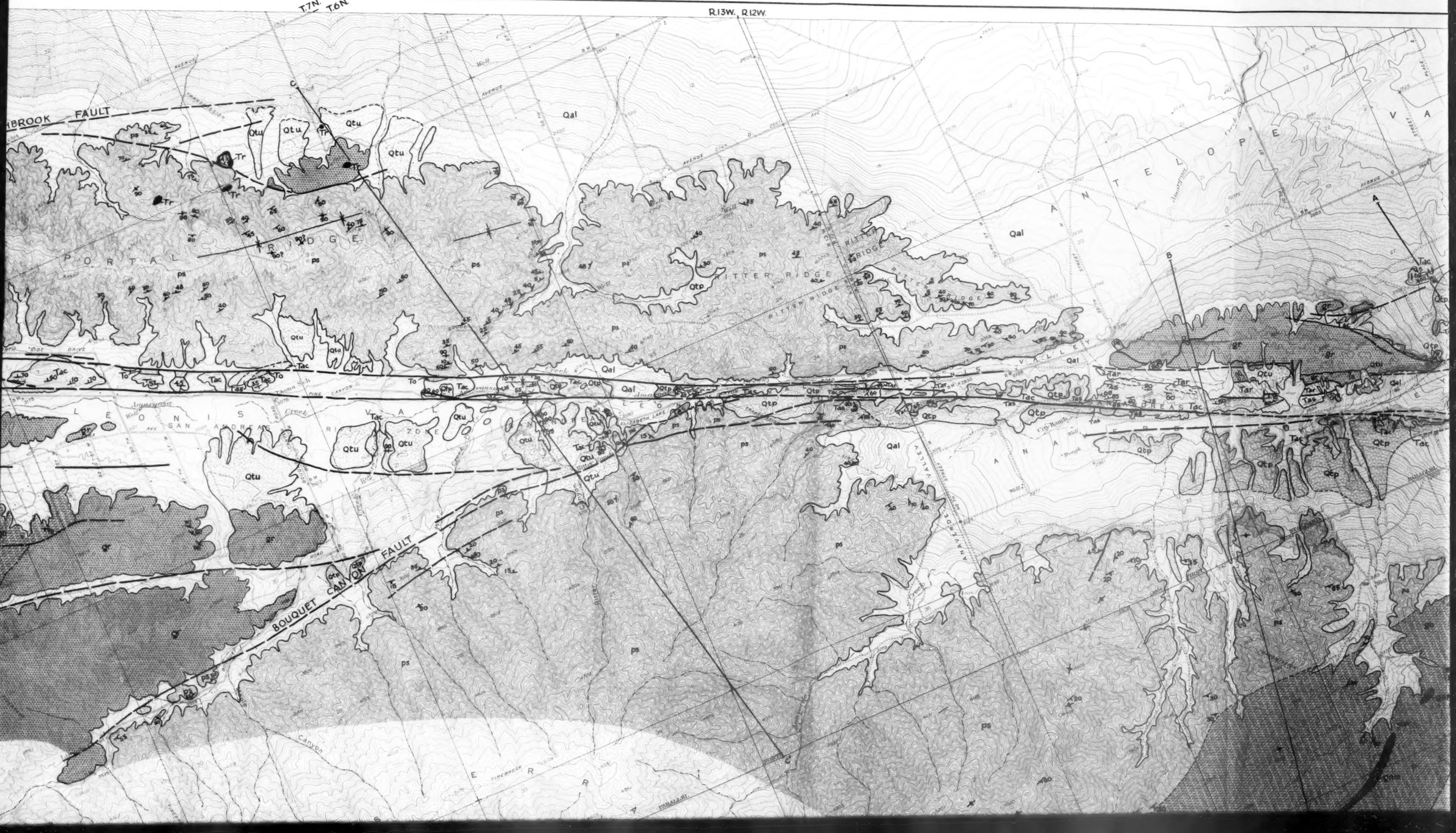
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PORTAL RIDGE
RITTER RIDGE
ANTELOPE VALLEY
LEONIS VALLEY
SAN ANDREAS FAULT
BOUQUET CANYON FAULT
Creek
Canyon
A
B
C

R.13W. R.12W.

T.5N.

R.13W. R.12W.

EXPLANATION

INTRUSIVE ROCKS

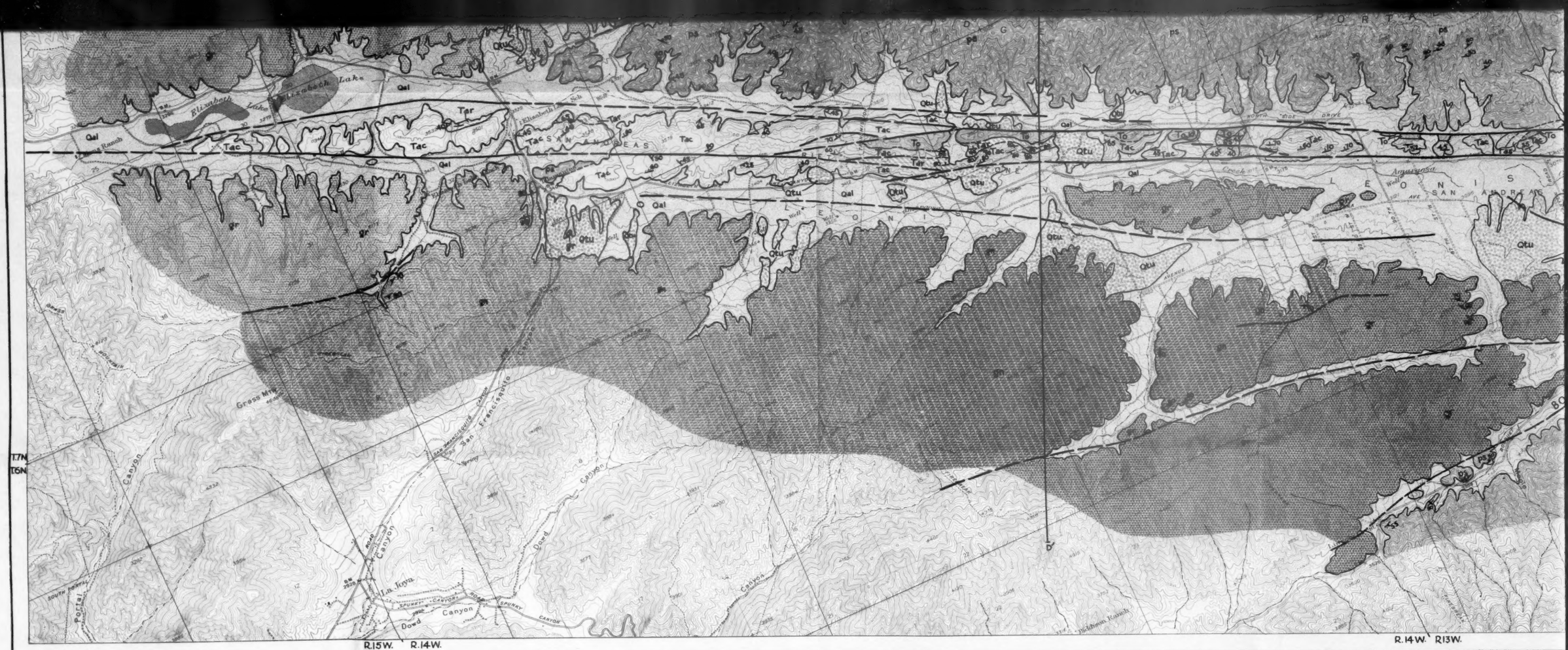
gr

Granite

(Including pegmatitic, aplitic, and dioritic phases, also mafic schist and gneissic zones)

Vertical scale = horizontal scale

ALTITUDE IN FEET ABOVE SEA LEVEL


Elizabeth Lake
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Portal Ranch
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San Francisquito Canyon
Dowd Canyon
La Veta
Spunky Canyon
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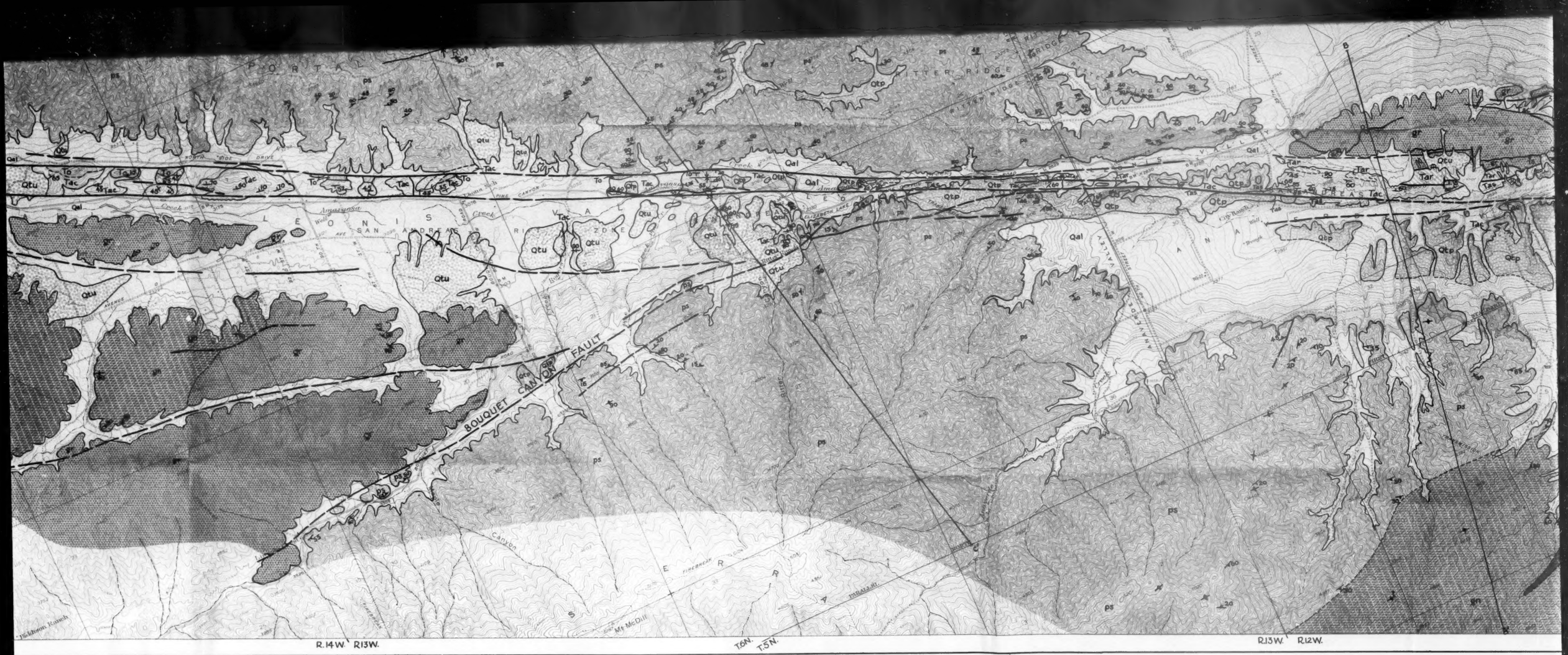
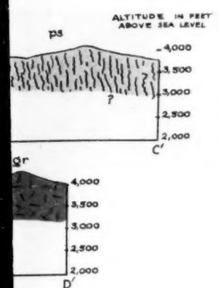




EXPLANATION

BEDDED ROCKS

- QUATERNARY
  - RECENT
    - Qal — Alluvium
  - PLEISTOCENE
    - Qtu — Undifferentiated terrace deposits
    - Qtp — Terrace deposits *(Composed of Pelona schist detritus)*
- TERTIARY
  - PLIOCENE
    - Tas, Tac, Tar — Anaverde formation *(Tas - shale; Tac - conglomeratic arkose; Tar - red conglomeratic arkose)*
  - MIOCENE?
    - Tr — Rosamond formation *(Acid flows and tuffs)*
  - OLIGOCENE?
    - Tvt, Tv — Vasquez formation *(Basalt and andesite flows and corresponding agglomerates and tuffs - Tvt)*
    - To — Old arkose *(Age uncertain)*
- PRE-CAMBRIAN?
  - ps — Pelona schist *(Principally meta-sediments, but including several intrusives and numerous quartz veins)*

STRUCTURE OF A PORTION OF THE SAN ANDREAS RIFT

GEOLOGY BY R. E. WALLACE 1941-1942

BASE MAP COMPILED FROM U. S. GEOLOGICAL SURVEY TOPOGRAPHIC MAPS

1 0 1 Miles

Scale

Contour Intervals 5 and 25 feet

N 16° W

EXPLANATION

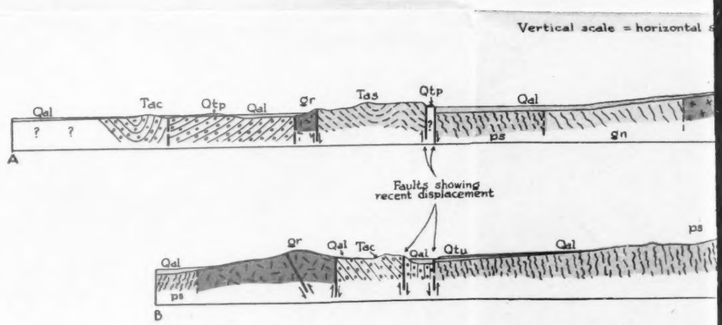
INTRUSIVE ROCKS

- PRE-TERTIARY
  - gr — Granite *(Including pegmatitic, aplitic, and dioritic phases, also mafic schist and gneissic zones)*
  - d — Diorite *(Including granitic phases)*
  - gn — Gneissic complex *(Biotite schists injected by diorite and granite, extremely distorted and metamorphosed. Mafic zone - gnm)*

SYMBOLS

- Contact
- Contact location uncertain
- 65° Fault U D *(Showing dip, upthrown and downthrow sides)*
- Fault location uncertain
- Anticline *(Axial plane vertical, axial plane inclined)*
- Syncline *(Axial plane vertical, axial plane inclined)*
- 30° Dips and strikes of bedded rocks
- 40° Dips and strikes of foliated rocks

Vertical scale = horizontal



EXPLANATION

INTRUSIVE ROCKS

gr — Granite
(Including pegmatitic, aplitic, and dioritic phases, also mafic schist and gneissic zones)

d — Diorite
(Including granitic phases)

gnm / gn — Gneissic complex
(Biotite schists injected by diorite and granite, extremely distorted and metamorphosed. Mafic zone - gnm)

PRE-TERTIARY

SYMBOLS

Contact

Contact location uncertain

65° Fault U D
(Showing dip, upthrown and downthrow sides)

Fault location uncertain

Anticline
(Axial plane vertical, axial plane inclined)

Syncline
(Axial plane vertical, axial plane inclined)

30° Dips and strikes of bedded rocks

40° Dips and strikes of foliated rocks

IFT

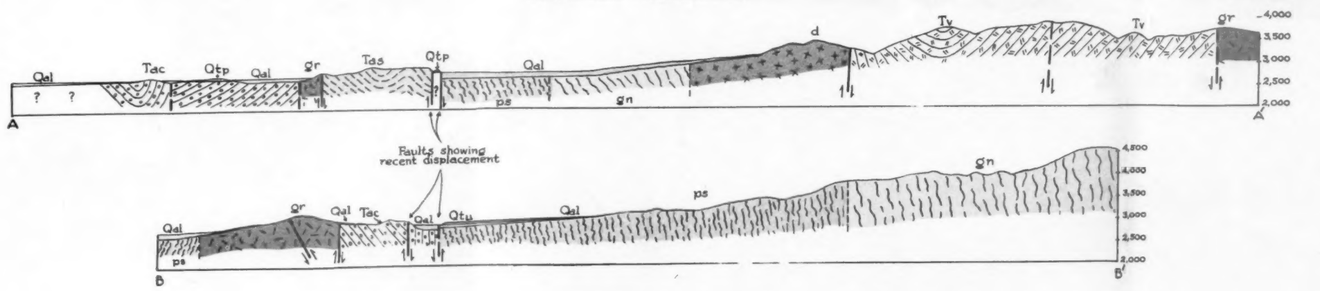
2

TOPOGRAPHIC MAPS

1 Miles

Vertical scale = horizontal scale

ALTITUDE IN FEET ABOVE SEA LEVEL



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STRUCTURE OF A PORTION OF THE SAN ANDREAS RIFT IN SOUTHERN CALIFORNIA

BY ROBERT E. WALLACE

CONTENTS

	Page
Abstract	782
Introduction	782
San Andreas rift problem	782
Location and size of area	783
Previous geologic work in the area	783
Acknowledgements	784
Geologic units	785
Generalized section	785
Pelona schist	786
Gneissic complex	787
Diorite	788
Granite	788
Old arkose	789
Vasquez volcanics	789
Rosamond volcanics	789
Anaverde formation	790
Terrace deposits	792
Structural geology	792
General statement	792
Recent faulting	792
Other structures in the rift trough	795
Other structures associated with the rift	797
Displacements in the San Andreas rift area	799
Rate of displacement	799
Geomorphic evidence	799
Structural and stratigraphic evidence	801
Antiquity of San Andreas rift	803
Mechanics of recent displacement	803
Summary	804
References cited	805

ILLUSTRATIONS

Figure	Page
1.—Index map of part of southern California	784
2.—Drainage map of the San Andreas rift and upper Santa Clara River area	791
3.—Diagrammatic sketch showing possible formation of center-trough ridge	796
Plate	**Facing page**
1.—Structure of a portion of the San Andreas rift in southern California	781
2.—Trough and recent scarp of San Andreas rift	792
3.—Aerial view of the San Andreas rift	793

ABSTRACT

An area approximately 20 miles long and 5 miles wide along the San Andreas rift in southern California between Palmdale and Elizabeth Lake was mapped. The area includes a portion of the Mohave Desert and two ridge zones separated by a trough along the rift.

The ridge zones are principally old crystalline rocks; the trough is underlain by a long, narrow strip of sediments which outcrop in a low "center-trough" ridge. Most of these sediments are part of the Anaverde formation which, on the basis of paleobotanical evidence, is believed to be transitional between lower and middle Pliocene. Evidence indicating the present relation of this long strip of sediments to the crystalline rocks on either side is conflicting. The older rocks of the ridge zones are in bands striking east or slightly north of east. This trend is cut by the rift which, in this area, has a strike of approximately N. 65° W.

Offset of terrace deposits suggests horizontal displacement along the rift of 5–6 miles since the Pleistocene. Streams are offset as much as 1½ miles. The north side of the rift has moved relatively southeastward. The situation of blocks of Pelona schist on opposite sides of the rift suggests displacement of 9 miles in the opposite direction to that indicated by recent faulting. This displacement is believed to be only apparent and possibly the result of differential vertical uplift of two separate Pelona schist blocks. Many and frequent vertical displacements of various blocks are indicated by elevated terrace deposits and old erosion surfaces.

There is no definite evidence that strike-slip displacement began earlier than the Pleistocene, although suggestions of greater age of the rift were found.

Strike-slip displacements similar to the San Andreas rift are apparently widespread in southern California suggesting that stresses have been exerted throughout a large area, possibly by drag from subcrustal currents.

INTRODUCTION

SAN ANDREAS RIFT PROBLEM

The San Andreas rift was first recognized as an important structural break by Lawson (1893, p. 151), and during the same decade Anderson (1899) and Schuyler (1896–1897, p. 711–712) described other parts of it.

The San Francisco earthquake and resulting fire of 1906 attracted the attention of geologists to the San Andreas rift and particularly to the strike-slip character of displacements. An investigation of the rift was immediately undertaken and a comprehensive report was prepared and published (Lawson, *et al.*, 1908). It was found that displacements had taken place over a distance of approximately 190 miles along the rift. Strike-slip was over 15 feet in many places and 21 feet in one place.

The next important contribution to the study of the rift (Noble, 1925–1926; 1927, p. 31) reported that in the portion of the rift north of the San Gabriel Mountains in southern California two masses of lithologically similar rocks at Cajon Pass and at Rock Creek on opposite sides of the rift are 24 miles apart, suggesting a possible 24-mile displacement. Noble also suggested that the San Gabriel and San Bernardino mountains may represent parts of the same mountain mass separated by strike-slip of about 75 miles. He emphasized, however, both in this paper and in a later publication (1931–1932, p. 355–363) that there is no proof of such great displacements.

After studying the Sunol fault of the Livermore region, Vickery (1925, p. 618–628) suggested that strike-slip of many miles occurred elsewhere on associated faults.

C. L. Gazin (unpublished report), while assisting Noble, found a possible strike-slip displacement of 5 or 6 miles suggested by the distribution of Pleistocene terrace deposits in the Pearland and Little Rock quadrangles. Gazin also noted stream-channel offsets suggesting strike-slip displacements of 1½ miles.

Noble (1925–1926), Willis (1938, p. 811), and others conjectured that the instigation of the rift might date back to early Tertiary. Reed and Hollister (1936, p. 84) suggested that the line of the rift might have been marked out even before late Jurassic time.

In contrast to the concept of the rift as an old feature with horizontal displacements of tens of miles, Taliaferro (1943, p. 160) found evidence in the central Coast Ranges suggesting that strike-slip has not been greater than 1 mile, that strike-slip is a very late feature cutting across structures and topography developed in the late Pliocene and mid-Pleistocene, and that the effects produced by the late Pleistocene and Recent movements along the rift are not comparable with those which resulted from Plio-Pleistocene diastrophism.

In addition, fossil vertebrate faunas from the Cajon beds at Cajon Pass and the Punch Bowl beds at Rock Creek, which were correlated by Noble, may prove to be of different ages (Chester Stock, oral communication).

The outstanding controversial questions regarding the San Andreas rift are:

1. Does the San Andreas rift date back to early Tertiary or pre-Tertiary time, or is the strike-slip an extremely recent feature cutting structural features formed by earlier displacement of different type?
2. Is the strike-slip of the rift many miles or less than a mile?
3. Does a different history of different portions of the rift account for the discrepancy of interpretations?

LOCATION AND SIZE OF AREA

The portion of the San Andreas rift mapped in the present investigation lies approximately 35 miles north of the business district of Los Angeles (Fig. 1). The area lies along the north flank of the Sierra Pelona Mountains, extends west from the Southern Pacific railway tracks, a few miles south of Palmdale, to approximately 1 mile west of Elizabeth Lake, and includes Portal Ridge, Ritter Ridge, and the adjacent portions of the part of the Mohave Desert known as Antelope Valley. Approximately 110 square miles are included in the map (Pl. 1).

PREVIOUS GEOLOGIC WORK IN THE AREA

The earliest published geologic account of the area was by Schuyler (1896–1897) who was interested principally in reservoir sites. Hershey (1902) described the Pelona schist and other metamorphic rocks in the area. Hill and associates made an unpublished reconnaissance map of the western portion in connection with a report on proposed reservoirs at Elizabeth Lake and Bouquet Canyon. Noble (1925–1926, 1927, 1931–1932) has been studying the San Andreas rift for over 20 years, chiefly in the area between Palmdale and Cajon Pass. He mapped the area first on the 1:62,500 topographic sheets of the Geological Survey. He is now remapping

it in much greater detail on the new large-scale topographic sheets that have become available for all of the rift that lies in Los Angeles County and on enlarged and topographically revised 1:62,500 sheets in San Bernardino County. Simpson (1934)

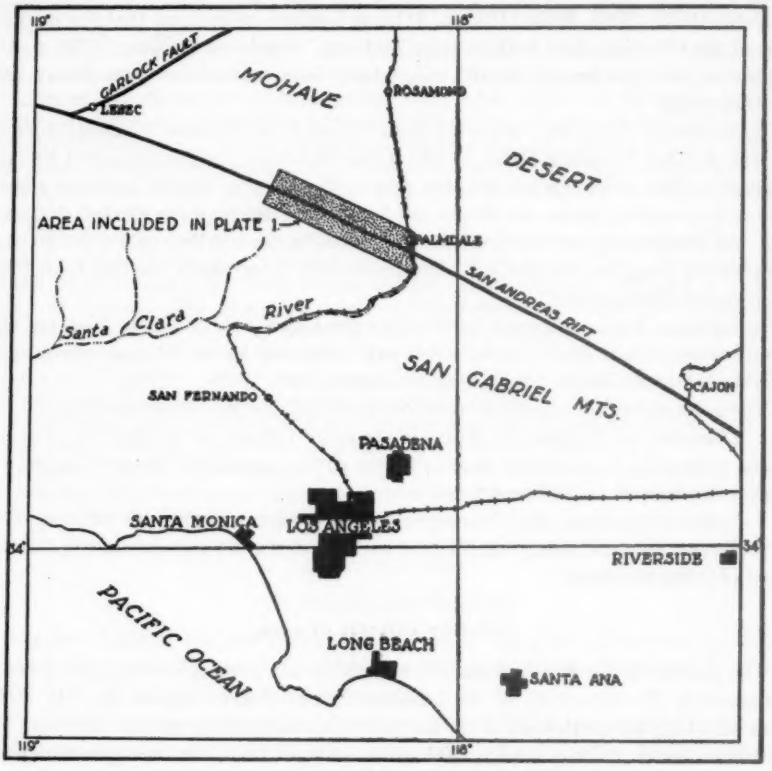


FIGURE 1.—*Index map of part of southern California*

mapped the entire Elizabeth Lake quadrangle on a reconnaissance scale. Gazin, working as Noble's assistant in 1931 and 1932, mapped the adjoining Pearland and Little Rock quadrangels in detail and also investigated the present area to some extent. An informal progress report on the Pearland and Little Rock quadrangles written by Gazin in 1932 was kindly made available to the writer by Noble.

Many investigators have studied adjacent areas. Among the more important papers are those of Kew (1924), Miller (1934, 1946), and Jahns (1940).

ACKNOWLEDGMENTS

The writer wishes to express his gratitude to Dr. J. P. Buwalda of the California Institute of Technology not only for suggesting this investigation of the San Andreas rift, but for his guidance and helpful criticism during the progress of the work. L. F. Noble of the U. S. Geological Survey offered numerous suggestions, and his help

and inspiration are greatly appreciated. Dr. D. I. Axelrod studied the plant fossils of the Anaverde beds. Finally, the writer wishes to express his indebtedness to the staff of the Division of the Geological Sciences of the California Institute of Technology for making available the equipment and facilities of the department during this study.

GEOLOGIC UNITS

GENERALIZED SECTION

A generalized section of the area is as follows:

	Sedimentary rocks	Thickness
		feet
Quaternary		
Alluvium	Unconsolidated sand and gravels	0–50+
	—Unconformity—	
Terraces, undifferentiated	Unconsolidated sand and gravels, essentially old alluvium	0–50+
Terraces	Pelona schist detritus; sand and gravels composed of mica schist and milky quartz cobbles and pebbles	0–50+
	—Unconformity—	
Tertiary		
Pliocene		
Anaverde formation	Gypsiferous shale	0–1000
	White to buff conglomerate and arkose	0–800
	Pink to reddish conglomerate and arkose	0–200
	—Unconformity—	
Miocene? or older		
Old arkose	Brecciated and pulverized arkose, dark red-brown	0–100?
	—Unconformity—	
	Igneous and metamorphic rocks	
Tertiary		
Miocene?		
Rosamond volcanics	Rhyolite tuffs and flows	0–300+
	—Unconformity—	
Oligocene?		
Vasquez volcanics	Basaltic and andesitic flows, tuffs, and breccias	0–3000+
	—Unconformity—	
Pre-Tertiary		
Granitic intrusive	Principally coarse-grained granite but also includes pegmatitic, aplitic, and dioritic phases	
Diorite intrusive	Principally diorite but locally includes granitic phases	
Gneissic complex	Biotite schists injected by diorites and granites, extremely distorted and metamorphosed	
Pre-Cambrian		
Pelona schist	Metamorphic complex, including metasediments, several intrusives, and quartz vein material	5000+

PELONA SCHIST

Pelona schist is the predominant rock formation in the area mapped. Rocks of this sequence crop out over most of the eastern portion of Portal Ridge and on Ritter Ridge and form a major portion of Sierra Pelona Ridge.

A great variety of metamorphic rocks are included in the Pelona schist sequence. For the most part they appear to represent metamorphosed sediments, although some of the rock units almost certainly were originally igneous.

The most common rock type is a mica-feldspar schist. The color grades from those rich in muscovite, which have a silvery sheen, to dark-brown schists in which biotite predominates.

Other schistose and gneissic types of rocks are present, although not so abundant as the mica schists. Schists containing abundant chlorite and amphiboles are common. A schist composed almost entirely of emerald-green actinolite crystals is very distinctive. Some of the crystals of actinolite are several inches long and an eighth of an inch or more in diameter. White fine-grained talc commonly accompanies the actinolite and is interstitial between the radiating actinolite needles. In some localities the talc predominates over the actinolite.

An augen gneiss forms several large areas of outcrop. The "eyes" of the gneiss are recrystallized feldspar grains, mostly albite and oligoclase. A gneiss also was found which was composed chiefly of clinozosite.

Quartzite beds a few feet thick are common in the schist sequence. In hand specimen the clastic characteristics are almost obliterated even though there is definite banding which is marked by fine sericite grains. In thin section the quartz grains show shadowy extinction indicating strain, and the borders of the grains are sutured. There is a definite preferred orientation of the quartz approximately paralleling the original bedding.

Marble beds are common in the Pelona schist sequence. They are finely crystalline and are white to gray. Graphite grains are common.

Quartz veins are abundant, forming pods and lenses parallel to schistosity in most places. The quartz veins range in width from a few inches to several tens of feet and consist of white "bull" quartz. In most cases no sulphides or other minerals are present in the quartz veins.

The presence of quartzites and limestone beds, clearly of sedimentary origin, interspersed throughout the mica schists suggests that the schist portions were originally sedimentary rocks. Hulin (1925, p. 26), analyzing similar relations in the Rand schist, commented:

> "The nature of these three original sediments, sandstone, shale and limestone, their purity and the character of their bedding as well as their areal extent and their thickness all leads to the belief that they represent a former series of marine sediments."

This statement is entirely applicable to the Pelona schist. The actinolite schists may represent intrusives, probably dikes, of mafic composition, although Hulin (1925, p. 27) suggests that the actinolite-rich zones in the Rand schist are the result of accumulations of basic volcanic tuffs in the original sedimentary series.

The schist sequence is folded, in some places tightly but commonly in relatively open folds. The folding appears to have followed metamorphism. Exposed sec-

tions of Pelona schist indicate a minimum thickness of over 5000 feet. It is very likely much more than that. Simpson (1934, p. 378) measured a thickness of more than 7500 feet.

The Pelona schist may be Archean. Hulin (1925, p. 23) suggested Archean age for the similar Rand schist, after noting that Paleozoic sediments in nearby mountain ranges are practically unaltered in comparison. The Placerita formation, described by Miller (1934, p. 4), which crops out near the western end of the San Gabriel Mountains in Placerita Canyon, may also be a correlative of the Pelona schist, although Miller did not believe so. Miller (1946, p. 457–542) has published the most comprehensive paper concerning the crystalline rocks of Southern California in which the correlation problems are discussed.

GNEISSIC COMPLEX

The gneissic complex crops out in two principal areas on the south side of the rift. One band of gneiss half a mile to 1½ miles wide extends southwest from near Palmdale reservoir to the south edge of the area mapped (secs. 4, 5, 6, 7, 8, T. 5 N., R. 12 W.). In general the zone is a complex of hornblende-feldspar-mica gneisses profusely intruded by granitic dikes which are themselves somewhat gneissic. Some of the dikes are about 100 feet wide. A few small limestone bodies suggest that this gneissic zone may represent a body of Pelona schist thoroughly injected by granites.

A dark, hornblende-quartz diorite band trends east-northeast parallel to the general strike of the gneissic zone and possibly represents a metamorphosed dike intruded into the gneiss.

The other area of gneiss borders the south side of Leonis Valley and extends from Grass Mountain (sec. 1, T. 6 N., R. 15 W.), east to approximately 1½ miles west of the Bouquet Canyon road in sec. 18, T. 6 N., R. 13 W. This zone is a mixture of gneissic, schistose, and igneous rock types. The typical gneiss to be seen on the road from San Francisquito Pass to Grass Mountain Lookout, consists of alternating light and dark layers, each from a fraction of an inch to several inches thick, in fairly regular bands which are twisted and folded into extremely complex structures; the dark bands are commonly colored by dark biotite, and the light bands are composed of feldspars and quartz. In some places the gneiss appears very similar to the Pelona schist but generally contains much more granitic material. Much of it may represent Pelona schist which has been thoroughly injected with granite and folded and contorted.

The age of this complex is doubtful. The San Gabriel formation in the San Gabriel Mountains (Miller, 1934, p. 49), apparently similar to the metamorphic-igneous complex described here, is a mixture of so-called Rubio metadiorite, Placerita metasediments, and an old granite, presumably the Echo granite. Miller (1934, p. 65) concluded:

> "The available evidence, then, points to the pre-Cambrian, and probably older pre-Cambrian age of the Placerita, Rubio diorite and meta-diorite, and Echo granite formations, as well as the greater part of the San Gabriel formation, but it must be admitted that the age is not absolutely proved."

At least some of the gneisses in the area mapped were probably formed by injection of the Pelona schist by younger granites.

DIORITE

An igneous mass which is slightly more basic than most of the granitoid rocks is located a mile south of Palmdale reservoir in sects. 8, 9, 10, T. 5 N., R. 12 W., and borders the block of Vasquez volcanics on the north and west sides. Although many parts of this block are true granite, most of the mass is diorite. The rock is typically brown to reddish and contains little or no quartz. On the southeast side the diorite is thrust up and over the Vasquez volcanics. The northwestern boundary was mapped as a gradational contact because no sharp contact was observed and it may or may not be a fault. In a road cut near the contact, gneiss is thoroughly injected with the granitic rocks, suggesting that the contact is essentially intrusive.

GRANITE

Several large blocks of igneous rocks of a granitic type are present in the area studied. The detailed relations of the gneisses and schists to the granite were not mapped. The granite blocks as mapped, therefore, represent areas which are predominantly granite, but which include gneisses and several types of metamorphic rocks.

North of the Hitchbrook fault in T. 7 N., R. 14 W. is the largest exposure of granite mapped. In hand specimens the granite is white, buff, or pinkish and grades in texture from coarse, almost pegmatitic, to very fine-graned. The granite block bordering the rift south of Elizabeth Lake is petrographically similar to that north of the rift.

Along the southern border of Leonis Valley a block of granitic rocks extends west of Bouquet Canyon approximately 2 miles. White to buff granite at the east end grades westward into gneissic rocks. Good exposures of the gneiss show that the granite is intimately injected into the gneiss as various-sized dikes and sills. In thin section the white granite commonly has less than 5 per cent mafic constituents.

North of the rift, approximately 2 miles west of Palmdale is a prominent ridge composed of granitic rocks. Most of the ridge is white to buff granite, but, along a reverse fault which dips south toward the rift, darker granite has been brought into contact with the white. At the western end of the ridge, at the mouth of Leonis Valley, the granite is apparently in intrusive contact with a small outcrop of Pelona schist. Although the contact is not well exposed, garnets in the Pelona schist indicate that it has been metamorphosed by the granite. Excluding the depositional contacts of alluvium, all other contacts of the block are faults. On the north side of the shale body just east of the mouth of Anaverde Valley small granite outcrops indicate that a thin body of granite represents an eastern extension of the large block. This stringer is apparently bordered by faults on both the north and south side.

Near Vincent on the west side of Soledad Pass in sec. 22, T. 5 N., R. 12 W. a block of granite is faulted against the Vasquez volcanics.

From the evidence in the area studied, the age of this granite can be limited only to post-Pelona schist time and pre-Vasquez time, but some apparently similar intrusives in the San Gabriel Mountains have been grouped by Miller (1934, p. 62–63) under the name of Lowe granodiorite.

OLD ARKOSE

A highly brecciated and mylonitelike sediment appears to be distinct from the Anaverde formation. Exposures of this rock along the north side of the center-trough ridge from approximately 1 mile east of the junction of the Bouquet Canyon road and the Pine Canyon-Elizabeth Lake road to a point approximately 4½ miles west of the junction are poor and commonly dark reddish brown. A few whole pebbles and cobbles were found which proved the sedimentary nature of the rock. Elsewhere the rock is so altered that no original structure could be distinguished. The color of the weathered outcrops appears similar to some outcrops of the Vasquez fanglomerates and red sandstones.

Near the western end of the old arkose outcrop area in sec. 2, T. 6 N., R. 14 W. the Anaverde sediments underlie the old arkose, but a comparison of the degree of metamorphism suffered by the two rock units suggests that the Anaverde formation is the younger of the two. It is believed, therefore, that the old arkose has been thrust over the Anaverde sediments.

VASQUEZ VOLCANICS

In an area of approximately 3 square miles in secs. 10, 11, 14, 15, T. 5 N., R. 12 W. bordering the west side of Soledad Pass, basalt and andesite flows and corresponding breccias and tuffs of the Vasquez formation crop out. The outcrops of the formation are confined to the south side of the rift. The Vasquez formation was originally named the Escondido formation by Hershey (1902, Pl. 1), but as that name was preoccupied Sharp (1935, p. 314) proposed the name Vasquez. Sharp measured a section of the Vasquez formation in the Ravenna quadrangle, which is just south of the area mapped, and found an approximate thickness of 9000 feet of fanglomerates and 4000 feet of basalts. Only the volcanics are exposed in the area mapped and are approximately 3000 feet thick.

The lavas range from massive, fine-grained basalts and andesites to amygdaloidal and vesicular masses. The amygdules are commonly filled with an agate deposit which weathers out of the basalt in the form of nodules, several inches in diameter. The flows are predominantly andesites and basalts. The surfaces of the flows were apparently very irregular so that tuffs deposited on their surfaces and later covered by other flows form irregular, lenticular bodies. The tuffs are of a crystal-lithic-vitric type, and have a distinctive green color.

The Vasquez formation has been determined to be various ages between Oligocene and middle Miocene (Simpson, 1934, p. 391; Miller, 1934, p. 70; Jahns, 1940, p. 170). The most recent evidence found by Jahns (1940, p. 170) suggests an age no younger than lower Miocene.

ROSAMOND VOLCANICS

On the north side of Portal Ridge in secs. 4, 5, 10, T. 6 N., R. 13 W. are several small patches of rhyolitic rocks which presumably represent the Rosamond series, widespread in the Mohave Desert, and over 4500 feet thick in the Fairmont Hills (Simpson, 1934, p. 400). Merriam (1919, p. 438–585) determined that the upper

portion of the Rosamond series is upper Miocene on the basis of vertebrate remains. The rhyolites in the area mapped by the writer are apparently in the lower part of the Rosamond (Simpson, 1934). Presumably the lower part of the section is also Miocene.

The largest outcrop of rhyolitic volcanics lies along the north side of the south branch of the Hitchbrook fault so that it is faulted against Pelona schist. The rather porous rock is white to buff in hand specimen. In thin section it is essentially a crystal-lithic tuff with fragments of rhyolite, quartz, and feldspar in a fine-grained groundmass. Some of the rock fragments and much of the groundmass is composed of fine-grained, chalcedonylike quartz. The whole mass appears to have been highly silicified.

On the easternmost granitic block in sec. 10, T. 6 N., R. 13 W. north of the south branch of the Hitchbrook fault, a small patch of rhyolitic debris was found which was angular and not water-worn, suggesting that it was the remnants of rhyolite in place. Two similar patches were found on the Pelona schist block south of the south branch of the Hitchbrook fault in secs. 4, 5, T. 6 N., R. 13 W. The rock is a porphyritic rhyolite with phenocrysts of quartz and albite-oligoclase in a fine-grained groundmass. The hand specimens are brown due to iron staining.

ANAVERDE FORMATION

The Anaverde formation is the principal Tertiary formation in the rift zone between Palmdale and Elizabeth Lake. Gazin (unpublished report) named the formation after Anaverde Valley where there are fairly good exposures of the sediments. Correlation is difficult because most contacts in the rift zone are fault contacts. The basal part of the series is a pink to reddish arkose in which granitic cobbles are present though not abundant. Noble (personal communication) reported that the basal arkose rests with depositonal contact on granite with pink feldspar at numerous places, all north of the trace of recent displacement. The red arkose is lenticular in places and grades laterally into the buff arkose which commonly overlies it. The red arkose appears to be approximately 200 feet thick in some localities.

Above the red arkose is a buff arkose that is the predominant unit of the formation. It is more conglomeratic than the red arkose and in some places contains boulders up to 1 foot in diameter although the common range is from pebbles a fraction of an inch in diameter to cobbles of 3 or 4 inches. The cobbles are almost all granite; in only a few places were gneissic cobbles found. In the upper part of this unit shale and siltstones are interbedded with the arkose. The arkose is well indurated commonly and in several localities weathers to form prominent ledges as do the conglomerates of the Cajon and Punch Bowl formations. The maximum thickness of Anaverde exposed in a single fault block is approximately 800 feet.

The top portion of the Anaverde formation is composed of brown to buff shales and siltstones. The shale is highly gypsiferous, and in the principal body of shale, just north of Palmdale reservoir, there has been small open-pit mining of gypsum. In one fault block the shale is approximately 1000 feet thick. It is not certain that this 1000 feet of shale is actually continuous above the arkose, and, since no fossil material which could be used for correlation has been found, the shale may represent

another formation. The shaly portions at the top of a continuous section of Anaverde arkose, however, suggest that the rest of the shale does represent the top portion of the formation.

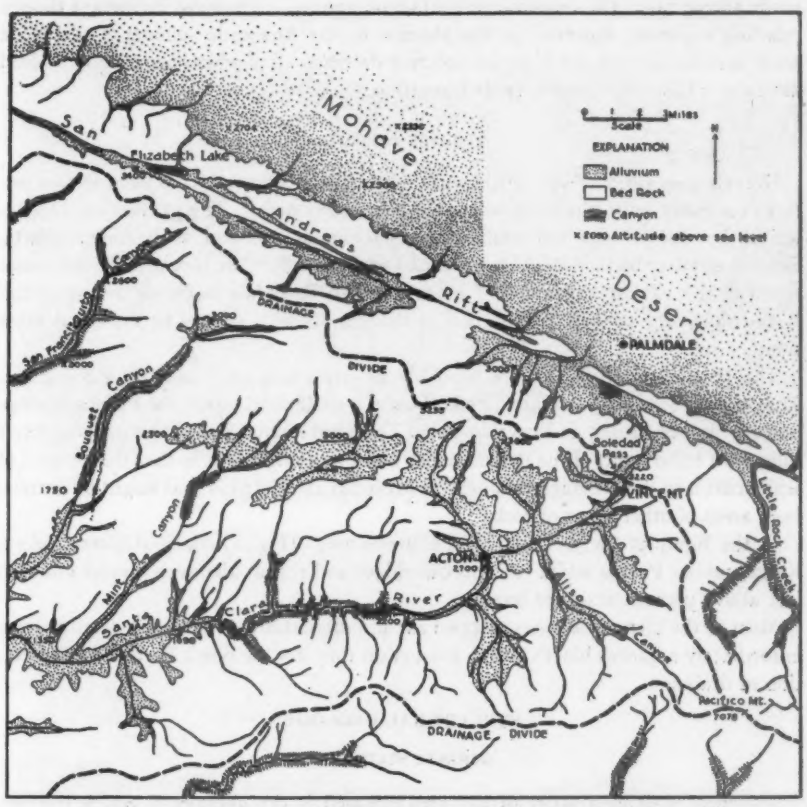


FIGURE 2.—*Drainage map of San Andreas rift and upper Santa Clara River area*

A small flora from silty beds near the top of the arkosic portion of the formation is the only fossil material found in the Anaverde formation. Fossil plant material was collected by both the author and Dr. Daniel I. Axelrod.

Nineteen fossil plants were provisionally identified by Axelrod, and he tentatively suggests that "the Anaverde flora may be intermediate between the Ricardo and Mount Eden floras in age, or that it is probably transitional lower to middle Pliocene".

The arkosic sediments with conglomeratic phases are typical deposits of a river in a semiarid climate which possibly had in part alluvial-fan characteristics. The shale and siltstone horizons represent flood-plain deposits, and the gypsiferous shale portions possibly represent playa lake deposits.

The source of the sediments is questionable. The granite cobbles of the arkose are almost identical with the granite north of Elizabeth Lake which is also typical of some of the other granites of the Mohave Desert. Noble has reported the Anaverde arkose resting in depositional contact on granite. The most important though puzzling evidence, however, is the absence in the Anaverde sediments of Pelona schist and Rosamond and Vasquez volcanic detritus, all of which presumably should have contributed to the Anaverde formation.

TERRACE DEPOSITS

Gravels composed almost entirely of Pelona schist detritus have been left as terraces on many fault blocks in the San Andreas rift area. The gravels are characterized by flat pebbles and cobbles of micaceous schists and white milky quartz. Since the source of the schist was almost certainly either the block of Pelona schist south of the rift approximately 2 miles south of Palmdale reservoir or the Portal Ridge block, the present distribution of the gravels with respect to the rift is most significant.

The Pelona schist gravels are probably of late Pleistocene to early Recent age. They overlie fossiliferous Harold beds (Gazin, unpublished report) of Pleistocene age but are faulted, tilted, and even folded. The predominance of schist in the gravels suggests a rapid uplift of the Pelona schist block as a single unit so that the amount of schist cast into the drainage channels far exceeded any detritus that might be derived from areas of other types of rocks.

In the Bouquet Canyon and Mint Canyon areas (Fig. 2) south of Sierra Pelona Ridge, similar Pelona schist terraces cap ridges and rest at altitudes several hundred feet above present drainage levels.

Most of the terrace deposits mapped as "undifferentiated" represent detritus from immediately adjacent blocks and in most cases they are the result of recent rejuvenation of drainage.

STRUCTURAL GEOLOGY

GENERAL STATEMENT

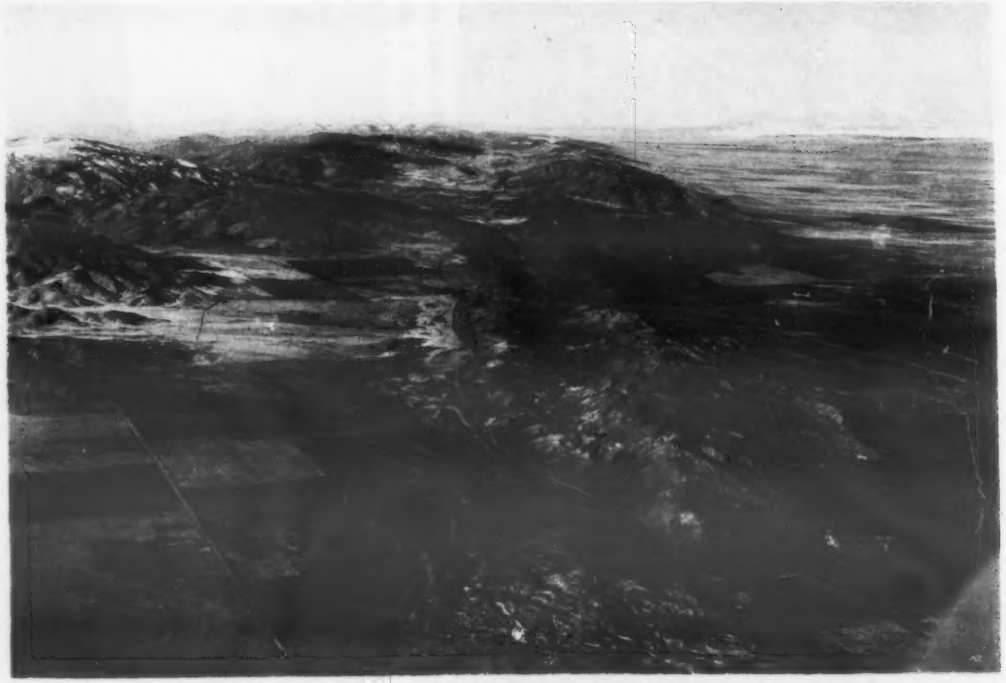
The most striking feature of the area mapped is the manner in which the San Andreas rift crosscuts the structural trend of the bands of older rocks. At least in recent times, the southwest side has moved relatively northwest. In the rift zone itself a long, narrow belt of sediments is bordered by faults which show evidence of recent displacements. The lines of recent movement strike approximately N. 65° W. Faults branching from the rift or cut by it strike approximately east and bound most of the blocks bordering the rift. There are several reverse faults and thrusts, most of which dip into the rift. Most of the folds trend east approximately parallel to the bands of older rocks. The blocks in and adjacent to the rift are tilted, distorted, and commonly brecciated.

RECENT FAULTING

Faults which are so recent that their scarp lines are essentially undissected are confined almost entirely to the topographic trough marking the line of the San

FIGURE 1. VIEW LOOKING SOUTHEAST FROM A POINT 4 MILES SOUTHEAST OF ELIZABETH LAKE
Showing typical recent fault scarp and sag pond.

FIGURE 2. VIEW LOOKING NORTHWEST ALONG THE LOWER PORTION OF LEONIS VALLEY THROUGH WHICH THE RIFT PASSES
A fault having recent displacement crosses the ridge spurs at left.



AERIAL VIEW OF SAN ANDREAS RIFT

Looking northwest from above Palmdale reservoir showing the Sierra Pelona Mountains on left, rift trough in center, Portal and Ritter Ridges right of center and the Mohave Desert on far right. Snow-covered Tehachapi Mountains are in right background. Photograph by Fairchild Aerial Surveys, Inc.

Andreas rift. Many of the associated faults such as those branching away from the trough zone and those bordering the Mohave Desert are also relatively recent but rarely marked by displacements in recent alluvium as are those in the trough zone. Commonly, however, the branch faults are followed by drainage which would rapidly destroy topographic expression of recent faulting. In some places the faults showing very recent activity are bordered on both sides by the same lithologic type of rock, suggesting small displacement. In general, however, the lines of recent movement appear also to represent much older lines of faulting along which repeated displacements have taken place. In a zone of faulting of this magnitude the most recent faults, and, therefore, the most easily traced topographically, are not necessarily the major break, which is actually a composite of many fractures along each of which movement has taken place at various times.

At the easternmost edge of the area the most prominent recent fault scarp crosses the Southern Pacific Railroad tracks approximately 2 miles south of Palmdale in sec. 2, T. 5 N., R. 12 W. The scarp forms the northern border of Palmdale reservoir. The north side of the fault is the upthrown side, and the escarpment, about 40 feet high, faces south. Anaverde arkose is exposed in the eroded scarp.

Approximately 1 mile northwest of the Southern Pacific tracks in sec. 34, T. 6 N., R. 12 W. most recent movement appears to have taken place along two branching faults a few hundred feet apart. The movement probably involved the ridge area as well as the depressed areas which seem to mark the faults. In several places in the ridge fault gouge cropping out from under the veneer of terrace deposit suggests that most of the ridge is gouge. Possibly compressional forces caused gouge behaving as a plastic to be squeezed upward forming a ridge.

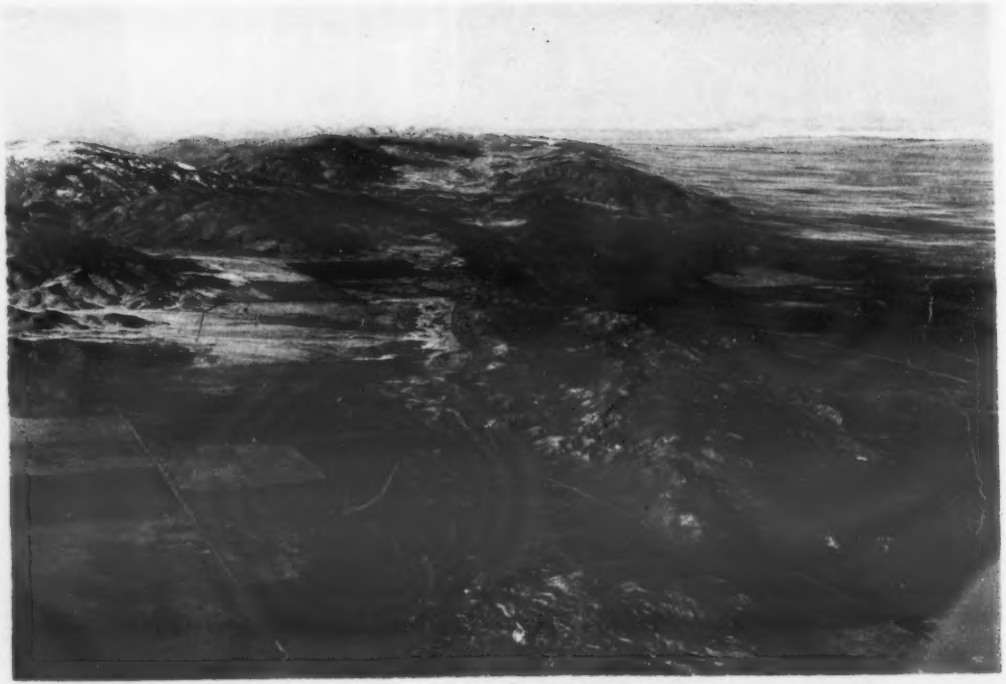
Approximately one third of a mile north of Lakeview in secs. 33, 34, T. 6 N., R. 12 W. where the line of recent displacements forms the south border of Anaverde Valley the fault scarp faces north. It is considerably more dissected than the one bordering Palmdale Reservoir and perhaps was formed during earlier movement along the fault. Anaverde arkose is exposed in the dissected scarp as in the scarp at Palmdale Reservoir. The presence of an upthrown block first on one side and then on the other side of a single fault line is a typical feature along the San Andreas rift.

To the northwest in secs. 32, 33, T. 6 N., R. 12 W. the fault is hidden by the alluvium of Anaverde Valley, but a more westerly trending fault branches off from the main fault. The scarp along this fault persists for approximately 1½ miles to a point in Anaverde Valley. The scarp faces north, and several sag ponds have been formed along its base, presumably due to subsidence of the block on the north side. For the most part the scarp is formed in gravels derived from the Pelona schist block. Excellent examples of terraces resulting from faulting have been formed.

West of the point where the main fault line crosses Anaverde Valley in secs. 29, 30, T. 6 N., R. 12 W., the fault trends diagonally at a very low angle across a ridge area. There it is marked by small sag ponds, pressure ridges, and valleys differentially eroded in gouge and shattered zones. The south side of the fault in this area was apparently upthrown so that the northern edge of the upper parts of Anaverde Valley were tilted to the south. This has caused a greatly decreased gradient in the streams draining from Sierra Pelona ridge into this portion of Anaverde Valley, and the lower sections of these valleys have been aggraded and "drowned" with alluvium.

, PL. 2

Looking northwest from above Palmdale reservoir showing the Sierra Pelona Mountains on left, rift trough in center, Portal and Ritter Ridges right of center and the Mohave Desert on far right. Snow-covered Tehachapi Mountains are in right background. Photograph by Fairchild Aerial Surveys, Inc.



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CE, PL. 3

Looking northwest from above Palmdale reservoir showing the Sierra Pelona Mountains on left, rift trough in center, Portal and Ritter Ridges right of center and the Mohave Desert on far right. Snow-covered Tehachapi Mountains are in right background. Photograph by Fairchild Aerial Surveys, Inc.

In the area where Amargosa Creek flows through a narrow canyon and where the two major blocks of Pelona schist are nearest together in sec. 24, T. 6 N., R. 13 W., the indistinct trace of the recent movement is apparently represented by a slight break in slope in the terraces underlain by Pelona schist south of the creek and the valley of Amargosa Creek itself. Anaverde arkose crops out on the south bank of Amargosa Creek in the western part of this constriction of Leonis Valley. From the west end of the narrow part of the valley to and beyond Elizabeth Lake, two faults branch from each other, separated by distances of from a few hundred feet to approximately one third of a mile. Along these faults is evidence of recent movement. The part between the two lines is for the most part a ridge, an example of the "center-trough ridge", but in many places there is no visible relief of the center block where it crosses alluvial areas. In such places aerial photographs will usually show fault traces; color of the surface differs along the fault due to variations of such features as moisture content of the soil and vegetation.

Approximately half a mile north of Ritter Ranch along the southern edge of sec. 15, T. 6 N., R. 13 W., the northernmost of the two faults is exposed in the cutbank of Amargosa Creek. The gouge zone along the fault is 8–10 inches wide; the fault is essentially vertical, with Anaverde arkose on the south side and Pelona schist detritus on the north side over the entire vertical range of the exposed fault. Minimum vertical displacement can be estimated at 50 feet. The south side is upthrown. This does not appear to be a case of apparent vertical displacement due to strike-slip.

A few hundred feet west of this location and also approximately half a mile west along the fault the Anaverde arkose is in contact with Pelona schist, but a thin veneer of alluvium covers the contact. In all other places alluvium or terrace material obscures any possible contact between Pelona schist or other rock bordering the trough and rocks of the center-trough ridge.

The center-trough ridge is the upthrown block along this fault, although Portal Ridge, topographically much higher and apparently a horst-like block, should be the upthrown side. Such evidence seems to indicate that the center-trough ridge and its bounding faults are essentially superficial features.

East of Elizabeth Lake schoolhouse in sec. 3, T. 6 N., R. 14 W. and sec. 33, T. 7 N., R. 14 W. the southern of the two recent faults extends diagonally across the center-trough ridge for approximately 3 miles. The fault is bounded on both sides by Anaverde arkose except for a short section just south of Elizabeth Lake School. Here a small wedge of what appears to be Pelona schist is present south of the fault. The Anaverde arkose south of the fault does not extend west of San Francisquito Canyon.

From Elizabeth Lake School to Munz Ranch at the west end of Elizabeth Lake in secs. 29, 30, 31, 32, T. 7 N., R. 14 W. the southern of the pair of recent faults is marked by scarplets and sag ponds so fresh and undissected as to suggest that the last movement was within historic times and certainly more recent than in any other part of the fault in the area mapped. At least some of this may have been the result of displacements during the Tejon Pass earthquake of 1857. The northern of the two faults is also marked by scarplets though not so clearly. West of Elizabeth

Lake the trace of the rift is confined to a relatively narrow valley and is bordered on both sides by granite. West of the area mapped at the head of Pine Canyon the fault trace crosses a relatively high mountain mass where the floor of the fault trough approaches an altitude of 4000 feet.

The ridge over 10 miles long in the center of the fault-trough zone is an upthrown block bordered by two faults of the San Andreas system, and is here called the "center-trough ridge".

Such an upthrown block in the center of a physiographically low zone suggests that the faults bordering the block do not have vertical displacements characteristic of the rift as a whole but are of secondary nature.

A possible origin of the center-trough ridge is diagrammed in Figure 3. The sequence of events is believed to be as follows:

A wide zone of fractured rock and fault gouge was eroded away and the edges of the adjoining blocks beveled off (Fig. 3A). Deposition of sediments on the floor of the valley took place (Fig. 3B). Lateral compressional components later squeezed plastic gouge upward in the direction of least resistance, and the central part of the sedimentary band above the gouge zone was faulted upward (Fig. 3C). Over a period of time erosion would cut down the trough and center-trough ridge more rapidly than the gouge material would be forced upward, so that the trough feature predominates over the ridge feature, and thus the crest of the center-trough ridge is lower than the crests of the two blocks on either side.

When fault gouge is squeezed upward the two blocks on either side move toward each other, causing buckling and thrust faulting in the sediments of the trough. The tendency of the gouge zone to become narrower and narrower due to the two blocks being forced together is compensated for by formation of additional gouge, the gouge zone adjusting to a width that can accommodate movements along the rift.

OTHER STRUCTURES IN THE RIFT TROUGH

Although the recent displacements in the rift and alluvial fill in the trough have obscured most older structures, a few faults and folds are well exposed. There are also many faults visible only on aerial photographs as alignments of difference in ground color. Many of these probably are relatively recent, as recent alluvium is affected, although they have no topographic expression.

Three faults parallel the rift in secs. 28 and 33, T. 6 N., R. 12 W., on the south side of the granite ridge approximately 2 miles west of Palmdale and north of the line of recent displacement. They are offset, however, by cross faults striking northeast. They may not have been involved in strike-slip displacements, or, if they were, the movement preceded the cross faulting.

In sec. 15, T. 6 N., R. 13 W. approximately half a mile north of Ritter Ranch, Amargosa Creek cuts across the center-trough ridge exposing a cross section of the ridge. The principal rock exposed is pinkish to buff Anaverde arkose which is folded into an asymmetric syncline paralleling the rift. On the north flank of the syncline strata dip 10°–20° S. and on the south flank 45°–55° N. A mass of Anaverde shale thrust over the south flank of the syncline in the arkose causes drag in the beds near the thrust plane.

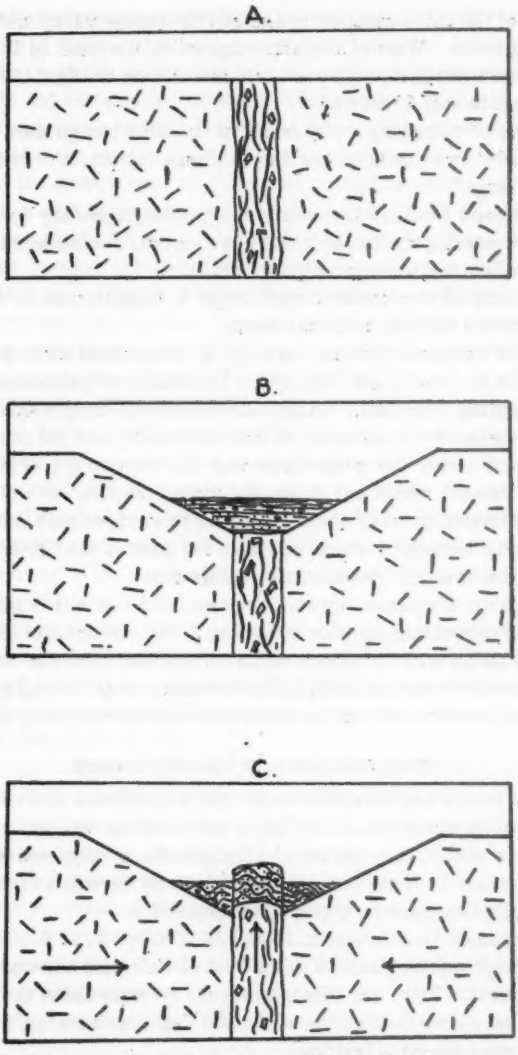

FIGURE 3.—*Diagrammatic sketch showing possible formation of center-trough ridge*
See text for explanation.

On the north side of the center-trough ridge from a point approximately 1 mile east to a point approximately 5 miles west of Leona School (sec. 16, T. 6 N., R. 13 W.), the old arkose is thrust over Anaverde arkose; in several places the thrust plane is nearly horizontal.

The traces of several other faults can be identified in the fault trough south of the

principal line of recent movement. Many portions of these can be seen more easily on aerial photographs than from the ground. In secs. 17, 21, 22, T. 6 N., R. 13 W. one fault extends from just south of the Elizabeth Lake-Pine Canyon road approximately half a mile west of the Bouquet Canyon road to the vicinity of Ritter Ranch. Over most of its trace it is not very distinct as observed from the ground, but aerial photographs establish its location without question. In secs. 17 and 18, T. 6 N., R. 13 W. another fault trace, approximately half a mile south of the Elizabeth Lake-Pine Canyon road between 87th Street West and 95th Street West, was identified only by means of aerial photographs. In line with this fault, and possibly representing a western extension of it, a fault is probably present but covered with alluvium.

OTHER STRUCTURES ASSOCIATED WITH THE RIFT

There are many structures in the mountain blocks on either side of the trough zone. Some are almost certainly associated with the rift structures; others may represent older structures formed at a different time and by different stresses.

In the area of the Vasquez volcanics west of Soledad Pass and approximately 1 mile south of Palmdale reservoir several reverse faults dip toward the line of recent displacement along the San Andreas rift. Along a fault which forms the northern boundary of the Vasquez volcanics a dioritic and granitic block has been thrust to the south up and over the volcanics. For approximately 2 miles the fault strikes nearly east and varies in dip between 60° north and vertical. At its west end the fault swings in a sharp right angle turn to a south strike and has a westerly dip of approximately 45°. Slickensides in the thrust-fault zone indicate that movement was parallel to the dip.

If the active force were directed from the southeast the thrust would be considered an underthrust. Another eastward-trending fault is entirely within the Vasquez volcanics and also has the north side upthrown. Between these two parallel faults the Vasquez volcanics are folded into an east-trending asymmetric syncline with axial plane dipping north. The fold is probably genetically associated with the faulting. The volcanics near the northward-striking thrust are also dragged and tilted to the east.

The south boundary of the Vasquez volcanics is also a nearly vertical fault. Presumably the granite which is present south of the fault has been faulted upward, since it is older than the volcanics. The Vasquez volcanics are broken along several other faults, none very distinct, as there is no contrast of rock types on either side to mark their traces. One fault can be traced over a distance of about 1 mile and trends N. 30° W. The northeast side appears to be the downthrown side. East of the central portion of this fault is a body of basalt which rests on top of vertically dipping, truncated beds of other volcanics. Whether the basalt is a flow on an erosion surface or a thrust block underlain by a horizontal thrust could not be ascertained. It seems more likely that the relationship is a result of faulting, because the rocks are all apparently of the Vasquez formation and faults are so numerous.

Another reverse fault dips 60° S.W. in secs. 28 and 29, T. 6 N., R. 12 W. on the north side of the line of recent movement in the west end of the granitic ridge 2 to 3 miles due west of Palmdale. In one good exposure of the fault plane, slickensides

indicate that the south side has been moved relatively upward. There is no indication of recent activity. The fault may continue northwest along the southern edge of Ritter Ridge as there is evidence in the Pelona schist of a fault of similar orientation. Slickensides are present on many rock faces of Pelona schist, and the truncated spurs facing Leonis Valley suggest faulting. Evidence supplied by slickensides must be used with caution because they are common on numerous fractures throughout most of the rocks, and practically all the structural blocks have been so thoroughly shattered that joint planes along which slight movement has taken place are present everywhere.

The Bouquet Canyon fault, the principal fault branching off on the south side of the rift, forms the northern boundary of the block of Pelona schist south of the rift. At Lincoln Crest, the divide at the head of Bouquet Canyon, the fault trace is marked by a distinct saddle. Half a mile to a mile east of this divide the Pelona schist block and the granitic block are in contact along the fault. A broad zone of shattered rock and gouge represents the fault. Undoubtedly the fault has determined the location of the upper portion of Bouquet Canyon and the valley east of Lincoln Crest. To the east the fault follows the north base of the Pelona schist block but for the most part is covered by alluvium. At one place a sharp break in slope of a spur of the Pelona schist block probably marks a branch, perhaps the major branch, of the fault. In the vicinity of Ritter Ranch (sec. 22, T. 6 N., R. 13 W.) the trace is covered by Quaternary terrace material, and to the east traces of several faults are present where the Bouquet Canyon fault should join the main zone of recent movement along the San Andreas rift. Approximately 2 miles from the San Andreas trough zone the Bouquet Canyon fault swings so that it trends slightly north of east, and at the head of Bouquet Canyon the strike of the fault is approximately N. 70° E. The age of first displacement along the Bouquet Canyon fault can be suggested only by the fact that Pelona schist detritus is present first in the Miocene Mint Canyon formation exposed in the Santa Clara Valley south of this area, but the greatest flood of schist detritus apparently was in the late Pleistocene and Recent.

The other large fault branching at a low angle from the San Andreas rift approaches and probably joins the rift west of Elizabeth Lake School (sec. 33, T. 7 N., R. 14 W.). It is marked by a prominent valley and strikes across Portal Ridge at approximately N. 80° E. passing through Hitchbrook Ranch. This fault, therefore, is referred to as the Hitchbrook fault. As with the Bouquet Canyon fault, Pelona schist forms the block on the south and granite the block to the north. On the north side of Portal Ridge the fault swings more nearly parallel to the strike of the rift, and there brings rhyolites of the Rosamond formation as well as granite into contact with Pelona schist. At the contact of the rhyolite and the schist the fault appears to dip steeply to the north. The north dip of this part of the fault suggests that it may represent an example of a tertiary thrust as described by Willis (1938, p. 802).

The principal line along which Portal Ridge has been raised above the Mohave Desert lies probably 1 mile or more north of Portal Ridge and approximately parallels the trend of the ridge. No physiographic expression was found to indicate the exact location of the fault, but the rejuvenated valleys on the north slope and terraces on top of the ridge are evidence of the uplift of the ridge.

DISPLACEMENTS IN THE SAN ANDREAS RIFT AREA

Rate of displacement.—Since the recognition of the strike-slip displacement at the time of the San Francisco earthquake of April 18, 1906, investigators have reported several strike-slip displacements along other portions of the San Andreas rift.

Recorded displacements give some idea of the rate at which movement is taking place along the line of the San Andreas rift. During the San Francisco earthquake of 1906 accurately measured strike-slip displacement ranged between 2 and 15 feet (Lawson, 1908, p. 133) (maximum of 21 feet was reported from one locality) over a distance of approximately 190 miles. In 1857 in the Tejon Pass region, displacements of over 30 feet took place, and the effects of the quake were visible for a distance of over 200 miles along the rift. During the earthquake of 1868 in the San Francisco area, strike-slip displacements ranged up to 10 feet (Lawson, 1908, p. 133). In the recent earthquake of 1940 in the Imperial Valley area (Ulrich, 1941, p. 13–31; Buwalda, personal communication) displacements averaged approximately 3 or 4 feet over a length of 40 miles. From these few cases a suggestion of the order of magnitude of displacement and rate of displacement is possible. A conservative estimate is an average of 5 feet displacement over a 100-mile length of the fault each 50 years. With that frequency and amount of displacement the rocks on opposite sides of the approximately 700-mile length of fault in California would be offset 5 feet in 350 years. A mile displacement would, therefore, accumulate in a period of about 400,000 years. If the rift has been active since mid-Tertiary time or about 30,000,000 years, then approximately 75 miles offset would result. Extrapolation to such a degree, of course, is dangerous reasoning, but even so, it indicates the amount of displacement possible. It also suggests the extreme effects that might be produced in only a relatively short lifetime of fault activity.

Geomorphic evidence.—The principal evidence of past strike-slip displacement along the San Andreas rift is the offset of stream channels crossing the rift. Examples from all sections of the rift length have been described.

Gazin (unpublished report) described the offset of stream channels in the Pearland quadrangle. Little Rock Creek (Fig. 2), one of the best examples, shows a horizontal offset of over 1½ miles, the north side shifting relatively to the southeast as is typical on all parts of the rift.

Drainage from Sierra Pelona Mountains is diverted eastward as much as 2½ miles in Anaverde Valley suggesting eastward shift of the north side of the rift by that much. Very possibly, however, at least the westernmost tributary of Anaverde Creek (Pl. 1) flowed across what is now a low ridge and out to the desert through the mouth of Leonis Valley (sec. 29, T. 6 N., R. 12 W.). The granitic block north of the rift which lies across the trend of these channels may have been uplifted vertically across the drainage thus shifting the flow to the east. The ridge, however, appears to be much older than the stream channels, and consequently the pattern is more easily explained by lateral movement along the rift.

In the Leonis Valley area, drainage from the Sierra Pelona Mountains has been dammed for some time by Portal Ridge so that it must flow northwestward or southeastward parallel to the rift. The present mouth of Leonis Valley may represent the mouth of Bouquet Canyon drainage offset to the east. It may on the other hand

be the result of a stream cutting southward from the north side of Ritter Ridge and capturing the Leonis Valley drainage.

Drainage from Portal Ridge into Leonis Valley also suggests horizontal displacements. For about 1 mile west from Leonis School (sec. 16, T. 6 N., R. 13 W.) along the center-trough ridge, notches in the ridge represent old channels through which Ridge drainage from Portal Ridge flowed. On the map (Pl. 1) the notches do not match the valleys on Portal Ridge. If the center-trough ridge were shifted east approximately 250 yards the notches and valleys would match much better.

In sec. 24, T. 6 N., R. 13 W., 1⅝ miles west of the mouth of Leonis Valley a stream joins Leonis Valley from the south side, and its mouth is offset 500 yards east of the rest of its valley, suggesting horizontal shifting of that amount.

Vertical displacements are apparent from a study of the geomorphology. On the drainage map (Fig. 2) the divide between the Mohave Desert and the Santa Clara River is asymmetrical, and the presence of this divide so close to the rift suggests vertical displacements along the rift zone. This asymmetry extends to a point a few miles east of Vincent where the drainage divide swings south away from the Mohave Desert. This, of course, is the result of the uplifting of the San Gabriel block and development of youthful drainage on both north and south flanks.

In the eastern part of the area mapped, from Soledad Pass west to Bouquet Canyon Pass, broad alluvium-filled valleys are present on the south side of the Mohave Desert-Santa Clara River drainage divide very nearly to the divide itself, but on the north side are steep-sided canyons. This suggests vertical displacements along the rift, and a comparison of altitudes of alluvial fill on either side of the divide indicates that this displacement in recent times may have been about 400 to 500 feet.

Patches of water-worn gravels are present at altitudes of 400 feet higher than that of Palmdale reservoir basin in sec. 10 and 11, T. 5 N., R. 12 W. on the block of diorite and granite approximately 1 mile south of Palmdale reservoir.

The alluvium south of the divide (Fig. 2) represents an older cycle of deposition and is separated from deposits of the present cycle by canyon portions on each of the valleys.

Much of the recent vertical displacement has taken place along a line obscured by alluvium of the Mohave Desert, but the canyonlike valleys on the north side of Portal and Ritter ridges indicate this displacement. Elizabeth Lake Valley and Leonis Valley in the trough are approximately 400 to 700 feet above the Mohave Desert level.

Recent vertical displacement has also accounted for the shifting of drainage in several places. Near Vincent large tributaries of the Santa Clara River enter the main channel of the river and make an acute angle with the downstream portion of the river. This feature strongly suggests that the headwaters of the Santa Clara once flowed northward through Soledad Pass. Indeed, Soledad Pass is broad and alluviated and undoubtedly was formed by a relatively large drainage channel, although now no water flows through the pass. The U. S. Geological Survey topographic map of the Elizabeth Lake quadrangel mapped in 1915 shows drainage flowing through the pass. The drainage divide at the pass is so low that perhaps the drainage has shifted even since the map was made. The reversal of drainage of these

two tributaries is due to a combination of uplift along the rift zone, thus shifting the divide northward, and increased gradient of the Santa Clara River which is causing erosion of deep gullies and arroyos in the older alluvium.

Surfaces representing old periods of erosion are present on most of the mountain and ridge blocks. On the ridge area north of the rift are a series of erosion surfaces and terrace deposits. On the west end above Elizabeth Lake the old surface is well preserved and is approximately 3700–4000 feet above sea level or more than 1000 feet above near-by portions of the Mohave Desert.

Between Elizabeth Lake School and Goode Hill Road the erosion surface on Portal Ridge is also at an altitude of about 3700–4000 feet but is not as well preserved or extensive as west of Elizabeth Lake School. East of Goode Hill Road on Ritter Ridge, the altitude of the surface is approximately 3300 feet. Farther east at the east end of the area mapped an erosion surface on the block of Anaverde shale 1 mile south of Palmdale is only 100–200 feet above the Mohave Desert. Patches of terrace deposits composed of Pelona schist detritus on both sides of the mouth of Anaverde Creek probably represent the level at which aggradation was taking place while the ridges were being eroded to produce the present upland surfaces of low relief. The present levels of the erosion surfaces and terrace deposits indicate that recently the west end of the ridge zone has been elevated more than the east end—possibly 1000 feet in the western part and 200 feet in the eastern part. The shape of these erosion surfaces suggests that the ridge area existed much as at present during the erosion period which produced the surface.

Erosion surfaces in the mountains south of the rift are not so prominent as on the ridges north of the rift. West of Bouquet Canyon only a few surfaces of low relief on the mountain tops suggest an earlier period of erosion. The crest of Grass Mountain is a good example, but the block as a whole is in late youth or early maturity with deep valleys developed on the flanks. The Sierra Pelona block has a relatively flat top. On the north side in secs. 26, 27, 28, 29, T. 6 N., R. 13 W. an even slope is preserved but is being dissected slightly by recent rejuvenation of the drainage.

Dissection suggests that the Grass Mountain block has been subjected to more recent elevation than the main elevation of the Sierra Pelona block. The Sierra Pelona block is actually higher, however, indicating that the main elevation of the Sierra Pelona block was much greater than that of the Grass Mountain block. Such alternate uplifting of adjacent blocks along the rift appears to be typical.

Structural and stratigraphic evidence.—Structural and stratigraphic evidence of strike-slip displacement along the rift is contradictory, but the best evidence strongly suggests horizontal displacements of several miles in a direction conforming with that displayed in recent displacements.

Gravels composed almost entirely of Pelona schist detritus and almost certainly derived either from the Portal Ridge block or the Sierra Pelona block of schist (Pl. 1) were found by Gazin (unpublished report) on the north side of the rift 5 or 6 miles east of the nearest bedrock outcrops of Pelona schist on the south side of the rift.

Gazin also pointed out that the most easterly terraces of schist composition are apparently older and are covered by younger granitic gravels. It is hard to escape

the implication that the north side of the rift has moved 5 or 6 miles relatively eastward and that the granitic gravels overlying the schist gravels were deposited as the north side of the rift moved past areas south of the rift which were casting granitic debris. This distribution, of course, might be explained as fluvial distribution by streams flowing eastward along the trough zone as in Leonis Valley at present. If this had been the case, however, the other blocks east of the Pelona schist should have contributed to the gravel, and the gravels should not be so predominantly Pelona schist.

In the area studied a narrow band of Pliocene Anaverde sediments over 20 miles long is present in the fault-trough zone. The material of the sediments is chiefly detritus from a granitic mass, whereas the rock masses bordering the band of sediments, and from which the sediments would be expected to be derived, are principally schists, gneisses, and volcanics.

The situation indicates drastic structural changes since Pliocene time. Perhaps the blocks on either side had such a structure and were involved in vertical movements of such a nature that schist, gneiss, and volcanics are exposed at present where in Pliocene time granite was exposed. The flood of Pelona schist detritus which took place in Pleistocene time and formed terrace deposits supports such a concept of post-Pliocene vertical displacements exposing schist. Contradicting this are the presence of Pelona schist in the Miocene Mint Canyon formation, and the occurrence of Miocene Rosamond volcanics apparently in place on an erosion surface of Pelona schist in the Portal Ridge area.

The relation may have been brought about by strike-slip displacements of 20 miles or more, but granitic blocks are not now situated as would be necessary if this had happened. The granite of the west end of Portal Ridge and of the Pine Canyon district is petrologically similar to the granite material of the Anaverde conglomerate.

The sediments might have been deposited by a stream flowing southeast along the rift from a high granitic block in the Pine Canyon area. However, schist detritus should then be found in the sediments unless the areas of schist outcrops were of such low relief that erosion was slight.

It is anomalous that the blocks of Pelona schist on opposite sides of the rift are apparently offset by strike-slip displacements opposite that supposedly typical of the San Andreas rift. The west edge of the Pelona schist block bordering the rift on the south side is 9 miles east of the west edge of the Pelona schist block in Portal Ridge. Both contacts are fault contacts with Pelona schist against granite. Orientation of structures in the two schist blocks is the same, commonly striking east.

It is, however, probably incorrect to compare the edges of the block only at the rift. Actually the western limit of the Pelona schist block south of the rift is over 10 miles west of the western tip of the Pelona schist block north of the rift. That the Pelona schist was exposed principally by vertical movements in Pleistocene time is suggested by the flood of Pelona schist detritus forming terraces; the present pattern of outcrops may be principally the result of these vertical displacements.

Another suggestion of anomalous displacement is the distribution of Anaverde sediments with respect to the line of recent movement. The westernmost outcrop of Anaverde sediments on the south side of the principal fault showing recent displace-

ment is approximately 2½ miles east of the westernmost outcrop of Anaverde sediments north of the rift. This, however, can also be accounted for by vertical displacements.

Vertical displacements of the blocks are indicated by the many terrace deposits on the blocks of Portal Ridge as well as those of the Sierra Pelona Mountains. Many of the undifferentiated terrace deposits (Pl. 1) are derived largely from adjacent blocks and are the result of recent rejuvenation of streams by differential uplift in recent times.

ANTIQUITY OF THE SAN ANDREAS RIFT

No clear-cut evidence of great antiquity of strike-slip displacement was found in the area studied.

Nothing older than the offset of terrace deposits of probable late Pleistocene or early Recent age can be attributed with any certainty to strike-slip along the rift. This by no means limits the age of strike-slip movement to post late Pleistocene, but rather emphasizes the magnitude of structural changes, whether horizontal or vertical, that have taken place along the line of the rift. Vertical movements have taken place in Recent time and are undoubtedly genetically associated with the strike-slip movement. If the great disarrangement of the blocks along the rift are principally the result of older vertical displacements, no evidence was found to suggest that strike-slip displacements were not associated then as in the Recent period.

Where cross sections of the faults showing recent displacement are exposed, the gouge zones are only a few inches to 1 foot wide. In contrast bodies of gouge several tens and probably hundreds of feet wide are present in the rift trough. Such zones seem to represent lines of much greater displacement and age than the lines of recent displacement.

It is peculiar that practically no outcrops of Anaverde sediments are exposed outside the trough itself. This may very likely indicate that the trough and therefore the rift existed prior to the middle Pliocene and that the sediments were deposited in the trough.

The presence of Pelona schist detritus in the Miocene Mint Canyon formation indicates exposure of Pelona schist at that time, but the flood of Pelona schist detritus forming Pleistocene and Recent terraces suggests that the principal uplift of the Pelona schist blocks adjacent to the rift was in the Pleistocene and later.

The problem of antiquity involves a specific definition of the San Andreas rift. The lack of a specific definition has apparently caused misunderstandings of various writers' interpretations. There are two important characteristics of the rift, the San Andreas line and the San Andreas strike-slip displacement. In speaking of the antiquity of the rift careful distinction should be made between these two, because, although there were probably ancient periods of diastrophism along the San Andreas line, the strike-slip movements may have had a relatively recent beginning.

MECHANICS OF RECENT DISPLACEMENT

Considering the present disagreeing interpretations of time and space relations of the rift's various structural features, an analysis of the overall mechanics of the

forces causing the features is impossible, but the nature of the cause of the recent strike-slip displacement may be suggested.

In southern California, Buwalda (1936, p. 341) has described faults in widely separated areas which show strike-slip displacements similar to that found along the San Andreas rift. Furthermore, Gutenberg (1941, p. 263–302) has described seismographic evidence which indicates that many shocks from various points in southern California apparently accompanied displacements having movements similar to that displayed at the surface along the rift.

Such a wide distribution of faults having essentially the same type of displacement would suggest that stresses were applied subequally throughout a wide area. Four origination points of stress on a crustal block are possible; from above the crust, from below the crust, tangentially on the side of the crustal block, and from within the block itself. The equal distribution of stresses can be derived most readily either from below or from within the crust. If it has been accomplished from below the crustal block, drag from subcrustal currents may be responsible. Griggs (1939, p. 611–650) has postulated such subcrustal currents and has described a possible interpretation of wedgelike mountain masses by such a process. He dealt principally with situations in which two crustal blocks would be dragged toward each other. Strike-slip displacements might be the result of streamlike currents with different velocities at different points across their widths which would create local stresses at various points in the overlying plate and would be capable of producing strike-slip displacements. A surface of discontinuity or "rip line" as in water currents, where two currents of different speeds and/or directions of motion meet, might be the line of change which would cause a "master fault" such as the San Andreas rift. Possibly a current might be flowing past a relatively stable mass. In this case the Sierra Nevada and associated tectonic blocks might act as the relatively stable mass. Seismographic (Gutenberg, 1943, p. 473–498) evidence indicates that the Sierra Nevada Mountains have roots to greater depths than the adjacent crustal units in California.

SUMMARY

The principal features of the faults with recent displacement may be summarized as follows:

1. The recent faults are marked principally by relatively undissected scarplets, pressure ridges, small sag ponds, gouge zones, and variations in soil and vegetation.
2. Faults showing very recent displacements are confined almost entirely to the trough zone.
3. The faults with recent displacements are essentially superficial and are believed to be only representative of the general path of the rift as a whole. The broad trough area represents a more nearly accurate trace of the San Andreas rift than does any single fault line within the trough.
4. Vertical displacements along the lines of recent movements are not indicative of the general displacements along the rift zone.

Other important structures in the rift area include:

1. The "center-trough" ridge which is believed to be formed by uplift over a zone of gouge and breccia acting in a plastic fashion.
2. Older faults which are parallel to the rift but which are cut by cross faults.
3. Thrust faults which in general dip toward the rift or are nearly horizontal.
4. Folds which parallel the rift or in places trend approximately eastward diagonally to the rift trend.
5. Bands of Mesozoic and older rocks which trend eastward.

The principal suggestions regarding amount of displacement in the San Andreas rift area can be summarized as follows:

1. Strike-slip displacements have caused offsets of stream channels of 1½ miles and of terrace deposits possibly as much as 6 miles.
2. Along individual faults with recent displacement there are stream offsets of several hundred yards, but rocks of similar type are present adjacent to one another on opposite sides of these faults. Contrasting this, the rock masses bordering the rift as a whole cannot be matched with sufficient certainty to suggest the total amount of strike-slip displacements.
3. The distribution of Pelona schist outcrops on opposite sides of the rift and outcrops of the Anaverde formation suggest anomalous displacement. The anomalies may be the result of vertical displacement.
4. Many and frequent vertical displacements of various blocks are indicated by present positions of terrace deposits and old erosion surfaces.
5. The thin band of Anaverde sediments, over 20 miles long and a fraction of a mile wide, composed of detritus different from the blocks bordering it, suggests drastic structural changes since Pliocene time. This may be the result of horizontal displacement or vertical displacement along the rift or of a peculiar arrangement of drainage.
6. The rate of recent strike-slip displacement, if extrapolated into the past, suggests 1 mile displacement in approximately 400,000 years or 75 miles displacement since mid-Tertiary time (about 30,000,000 years).

Observations regarding the antiquity of the rift are as follows:

1. A differentiation between two San Andreas rift characteristics must be made; that is, between the San Andreas line and the strike-slip displacements along the rift.
2. No evidence was found to prove an age of the strike-slip displacement greater than Pleistocene, but many suggestions that the line was marked out earlier than Pleistocene were found.

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